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THE CHALCOCITE AND NATIVE COPPER TYPES OF ORE DEPOSITS.

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IN certain types of copper deposits the metal is present largely in the form of native copper or chalcocite, and chalcopyrite and pyrite, so common in most copper deposits, are absent or rare. Expressed in chemical terms this means relative richness in copper and paucity of iron and sulphur. The extreme of contrast in this regard is exemplified by a comparison of such deposits as those of Lake Superior, where the copper is mostly native, with deposits such as those of Rio Tinto, Spain, where the dominant copper mineral is the relatively lean copper sulphide, chalcopyrite, abundantly accompanied by pyrite. Between these extremes are the deposits of Corocoro, Bolivia, with both native copper and chalcocite abundant, the chalcocite replacements of limestone at Kennecott, Alaska, the deposits of Mansfeld, Germany, with bornite as their dominant copper mineral, the "red beds" type of chalcocite deposits, and finally, the chalcocite zones resulting from downward enrichment.

These types of deposits form the subjects of a large but scattered literature. The principles involved in the formation of one type—the chalcocite zones of downward enrichment—are now well understood, but with respect to the other types much diversity of opinion still exists. This diversity in opinion involves fundamental problems of ore genesis which it is the writer's purpose to discuss in this paper. Certain of these deposits are secondary,

owing their rich copper minerals to the modification (enrichment) of older deposits poorer in copper. Such deposits are common. Others—notably the native copper deposits—are comparatively rare and are in the main primary. Were these deposited from solutions of magmatic derivation? If so, did these solutions leave the magma relatively rich in copper and poor in iron and sulphur, or were they robbed of their iron and sulphur as they travelled from their source to the final seat of ore deposition?

Standards of Comparison.—Although consideration of the primary native copper deposits will be postponed to the second part of this paper, it is desirable at this point to set up certain standards with which the chalcocite and native copper ores may be compared. Of late years the genetic relationship of the majority of ore deposits to vulcanism has been widely accepted. Under this concept the majority of copper ores are either igneous rocks of unusual composition or are deposits in the country rocks from solutions emanating from crystallizing magmas. These classes of magmatic and hydrothermal deposits conjointly give much information as to the nature and proportions in which elements were present in the magma, the one indicating the components that remained in the magma, the other those expelled in solution. The proportions in which copper, iron and sulphur are present in such ores may be significantly compared with their proportions in ores whose genesis is attributed partly to other processes.

Magmatic copper-bearing deposits in the narrow sense of that term are exemplified by many of the Norwegian sulphide-bearing gabbros and related rocks described by Vogt,[1] by sulphide-bearing norites and pyroxenites in Connecticut described by Howe,[2] by a somewhat similar occurrence described by Bastin [3] from Maine, and by many other occurrences. In such rocks copper occurs characteristically as chalcopyrite (bornite being only exceptionally

[1] Vogt, J. H. L.: Zeit. für prakt. Geol., 1893, 125–143.

[2] Howe, Ernest: Sulphide-bearing Rocks from Litchfield, Conn. Econ. Geol., vol. 10, 1915, 330–347.

[3] Bastin, E. S.: A Pyrrhotitic Peridotite from Knox County, Maine. Jour. Geol., vol. 16, 1908, 124–138.

present) and is accompanied by pyrrhotite, rarely by pyrite, and commonly by pentlandite. It is clear that in the magmas from which they crystallized, iron and sulphur were sufficiently abundant not only to form chalcopyrite with all of the copper present, but to form also abundant pyrrhotite.

Many of the leaner sulphide-bearing rocks of the Sudbury copper and nickel producing district probably belong to the class of strictly magmatic deposits, and the great commercial ore bodies of that district were clearly of magmatic derivation, although mineralizers appear to have played a rôle in their formation. In these ores, copper occurs as chalcopyrite and is accompanied by abundant pyrrhotite, some pyrite and pentlandite.

The contact-metamorphic or pyrometosomatic type of copper deposit is well represented in North America by some of the primary protores or ores at Clifton and Bisbee, Arizona, and Cananea, Mexico. Deposits of this type are characterized by chalcopyrite as the dominant copper mineral associated with abundant pyrite, more rarely pyrrhotite, and commonly with magnetite or specularite. Minor amounts of bornite may be present. It is clear that the mineralizing solutions were rich in both iron and sulphur.

In most deposits clearly of hydrothermal origin, chalcopyrite is the prevailing copper mineral in the primary ore, and pyrite (or rarely pyrrhotite) is an abundant accompaniment. It is true that in some deposits, as at the Braden Mine, Chile, and at Butte, primary bornite is also present but not to the exclusion of chalcopyrite and pyrite. The sulpho-salts of copper—enargite, tetrahedrite, etc., common in some of these deposits, are richer in copper than is chalcopyrite, but these combinations of copper with arsenic or antimony and sulphur are not present to the exclusion of either chalcopyrite or pyrite.

It appears, therefore, that in those copper ores that are clearly magmatic and in the vast majority of copper ores deposited by solutions emanating from magmas, iron and sulphur are present in sufficient abundance to form chalcopyrite by combining with copper (although copper may also combine with arsenic or antimony) and that pyrite or pyrrhotite forms in addition. This

appears to be true regardless of whether the parent magma be acid or basic. This common mode of copper occurrence will, for the sake of brevity, be referred to hereafter as the "standard" mode of copper occurrence. With this commoner mode of copper occurrence the ore deposits that form the special subject of this paper are in marked contrast. In them chalcopyrite and pyrite are rare or absent, signifying that iron is absent or relatively rare and sulphur is much less abundant. Their dominant copper minerals are chalcocite and native copper and less commonly bornite or covellite. Let us consider first the best understood among these types.

THE CHALCOCITE ZONES OF DOWNWARD ENRICHMENT.

The processes of downward enrichment in copper deposits are now so well understood that they need be considered here only briefly as a basis for a fuller understanding of other ore types especially rich in copper.

If the primary ore was of the common or "standard" type carrying chalcopyrite and pyrite as its dominant metallic minerals, and if we recognize chalcocite as the end product of copper enrichment, then, reduced to its simplest terms, the process of downward enrichment if carried to completion consists in the removal of all of the iron and some of the copper and sulphur in solution and the redeposition of the remaining copper and sulphur as chalcocite- Cu_2S.

The separation of copper from iron in oxidation and downward enrichment may conceivably be due to (1) selective solution, (2) selective precipitation or (3) a combination of both processes. If the sulphides of iron in an ore oxidize at different rates from those of copper, a partial separation of the two metals will result. Experimental studies under controlled conditions, so far as they go, are in accord with field observation, which indicates that in pyritic ores containing bornite and chalcocite, the latter are likely to be first attacked. As to the relative susceptibilities of chalcopyrite and pyrite, field evidence indicates that chalcopyrite commonly yields first.

Whatever may be the effects of selective solution, it is clear that selective precipitation is of dominant importance. Without going into unnecessary details in regard to reactions that are well understood, it should be recalled that the principal products of oxidation of sulphide ores of copper are ferrous sulphate ($FeSO_4$) and cupric sulphate ($CuSO_4$). These two behave very differently under the conditions of the zone of oxidation and under the reducing conditions of the ground-water zone into which they descend. In the oxidized zone some of the ferrous sulphate is likely to alter to ferric sulphate and eventually to become fixed in the gossan as "limonite." The bulk of the iron, however, moves downward in sulphate solution into the ground-water zone where it is carried out of the ore body into the general ground-water circulation. Copper, except as it may be held in the oxidized zone as carbonate, silicate or crystallized sulphate, descends as $CuSO_4$ into the zone of precipitation where, in contrast to iron, it is largely—though not completely—reprecipitated as sulphides. The dominant mechanism of precipitation is the replacement of earlier sulphides.

If we accept the formula $Cu_2S.Fe_2S_3$ as best expressing the composition of chalcopyrite—the dominant primary copper mineral—then the net effect of the development of chalcocite (Cu_2S) as the end product of enrichment is the removal of iron and of sulphur in balance with it. Native copper is not a characteristic product of the zone of enrichment although, as is well known, it may develop in the zone of oxidation as a residue from the oxidation of the sulphur of chalcocite.

The separation of iron from copper in downward enrichment is therefore a consequence mainly of selective precipitation due to the absence of effective precipitating agents for the iron, with which the copper was originally associated in the primary ore.

THE RED BEDS TYPE OF COPPER DEPOSITS.

Though commonly of little economic consequence, the copper deposits associated with the Red Beds have attracted wide interest, and form the subject of a voluminous literature, the important items of which are listed in the Bibliography.[4]

[4] Numbers in parenthesis refer to bibliography at the end of this paper.

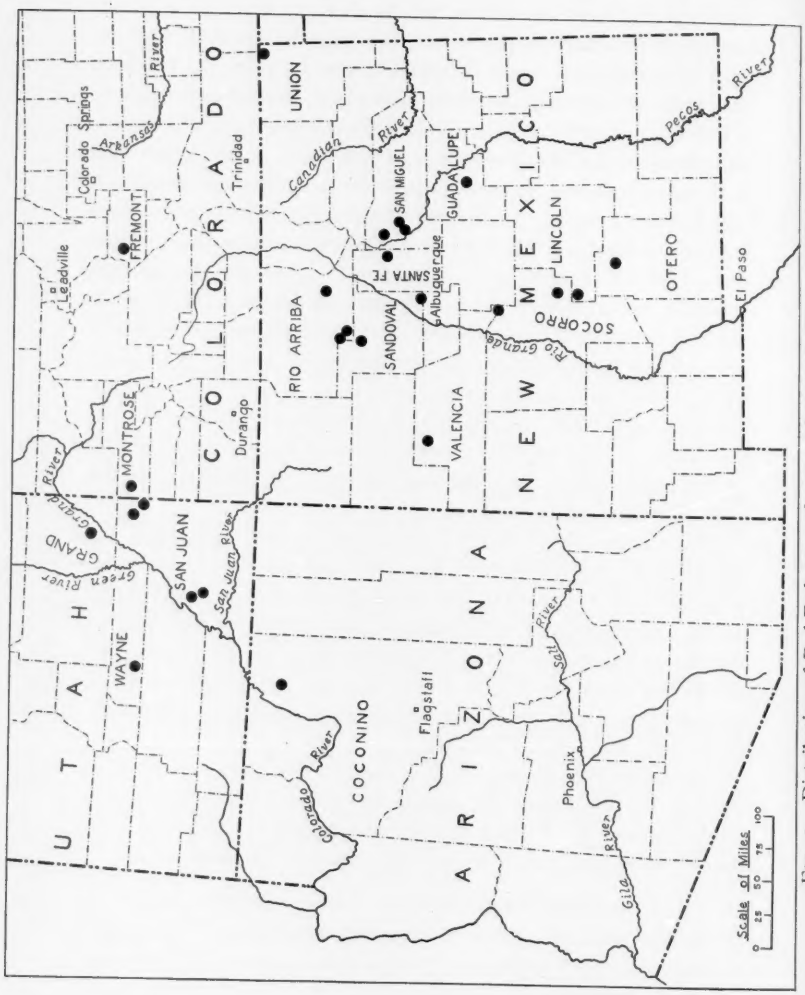

FIG. 1. Distribution of Red Beds type of copper deposits in Western United States.

Distribution.—In the United States deposits of this type occur widely in the states of Colorado, New Mexico, Arizona, Oklahoma and Texas, the principal occurrences that have been described being shown in Figs. 1 and 2. The most extensive foreign occurrences are on the plains or steppes west and southwest and southeast of the Urals in Russia at the localities shown in Fig. 3, but other occurrences are widely distributed over the

FIG. 1. Distribution of Red Beds type of copper deposits in Western United States.

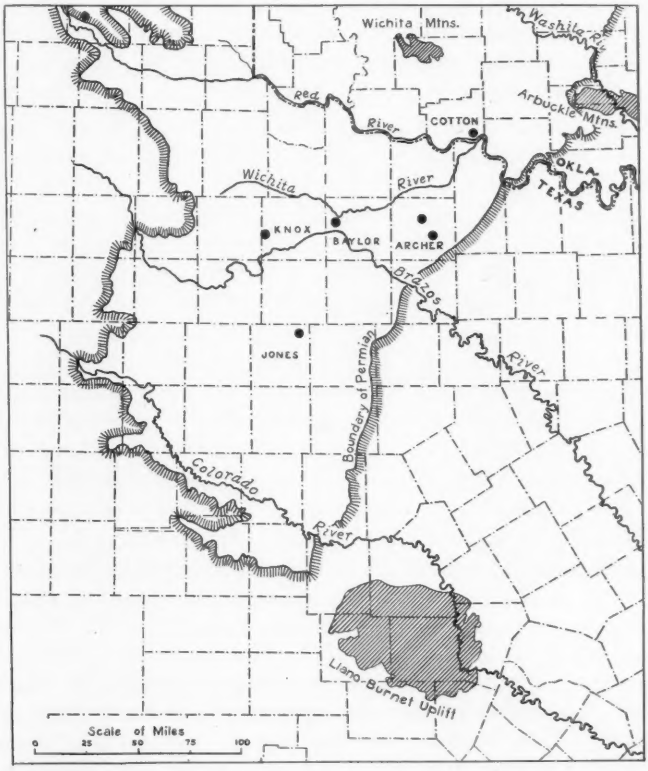

FIG. 2. Distribution of Red Beds type of copper deposits in Texas and Oklahoma in relation to ancient uplands and the boundaries of the Permian Red Beds.

world, in Germany, Czechoslovakia, France, Portuguese West Africa, Nova Scotia, New Brunswick, and in the province of Hon-peh in China.

Age of the Enclosing Rocks.—Deposits of this type occur in rocks of Pennsylvanian, Permian, Jurassic, Triassic, Cretaceous and possibly Tertiary age, although they are most widespread in the Permian.

Petrology of the Enclosing Rocks.—The enclosing rocks are without exception sediments; igneous rocks are characteristically absent from the locality and commonly from the entire region; so that it is impossible to attribute the ores to any nearby igneous

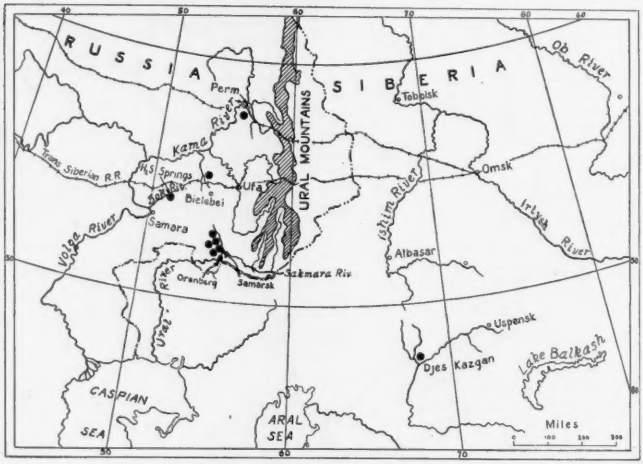

FIG. 3. Distribution of Red Beds type of copper deposits in Eastern Russia and Western Siberia.

sources. The prevailing rocks are sandstones, commonly arkosic and cross-bedded but with them some clay shales are usually associated. Fine conglomerates may also be present. The colors range from purplish to brown, red, pink, yellow, gray and white. Not uncommonly the matrix of the sandstone is calcareous and in a few localities thin interbedded limestones occur. In several

localities interbedded gypsum is present and, in some of the Russian localities, salt and anhydrite. Rarely, as at Hohenelbe in Czechoslovakia, bituminous shales occur close to the ore-bearing beds.

Plant Remains.—In about twenty-five of the forty or so consulted descriptions of ores of this type the association of plant remains with the ore—particularly the richer ore—was sufficiently striking to be commented upon by the observer.[5] It is unnecessary to cite these examples in detail but in some instances tree trunks up to 60 feet in length have been replaced and mats of vegetation are impregnated with copper minerals. Some of the replaced logs assay 47 per cent. copper. A few deposits are argentiferous. The occurrences indicate the frequency and intimacy of association of such ores with plant remains—an association that can hardly be fortuitous. It seems evident that organic matter when present has exerted an important influence in ore deposition, the precise nature of which will be considered later.

On the other hand, in the descriptions of many of these deposits, the occurrence of plant remains is not mentioned or they are recorded as only sparsely present. In some cases they were probably not recorded because their significance was not appreciated; in others, it seems clear that they are absent or rare.[6] It therefore seems probable that agencies other than direct contact with plant remains have been operative in copper precipitation. This problem will receive attention later.

Significance of Color.—The predominance of red color in the formations with which these ores are associated has led to their designation as the "Red Beds type" of copper deposit. When, however, the descriptions of occurrences are critically studied, a number of deposits conformable to type in all other respects are found not associated in any way with red beds and the deposits that are associated with prevailingly red formations are very commonly in their gray or white facies.[7]

[5] See Bibliography references: 5, p. 100; 10, p. 145; 14, p. 172; 19, p. 591; 24, p. 147; 26; 34, p. 154; 35, p. 142; 38; 39, p. 677; 42.

[6] See Bibliography references: 13; 19; 24; 27; 41; 43.

[7] See Bibliography references: 19, pp. 621, 614; 17; 26; 24, p. 118; 13; 31, p. 117; 20; 24, p. 144; 34, p. 144; 37, p. 193.

In summation: The available evidence indicates that the ores of this type have been deposited indiscriminately in sediments of various colors from white to pink, yellow, red and brown. Where the ores are associated preferentially with bleached facies of prevailingly red formations there is no evidence that this bleaching was a consequence or accompaniment of mineralization. On the contrary, it seems to be a consequence of the abundant presence of plant remains. These remains, however, may later have exerted an influence upon ore deposition.

Source of Copper of the Red Beds.—Since the ores under consideration are not associated with igneous rocks, they cannot well be deposits from solutions emanating from magmas and they have commonly been attributed to the destruction through weathering of older and more or less remote copper deposits usually of the "standard" type, carrying abundant chalcopyrite and pyrite and closely associated with igneous rocks. The conclusions of Murchison with respect to the deposits in the Permian beds of Eastern Russia are of especial interest in this connection, to quote:—

> That portion of the Permian strata which is cupriferous extends for a certain distance only to the west of the Ural chain (on the average from 265 to 330 miles). In all the Permian tracts more distant from these mountains no trace of the copper ore is to be found. These circumstances alone would naturally lead to the belief that the Ural Mountains had afforded the sources from whence the mineral matter proceeded. As this chain was, in remote periods, the seat of processes of intense metamorphism, during which copper veins were abundantly formed in the older Paleozoic rocks, we are naturally led to suppose that such operations may have had some connection with the deposit of the adjacent copper sands and marls.

The copper, Murchison states, was transported in solution and not mechanically.

The deposits near Hohenelbe and Wernersdorf in Czechoslovakia are contiguous to the crystalline area of the Risengebirge, which contains numerous copper deposits associated with igneous and metamorphic rocks (39, p. 685).

In the United States the pre-Cambrian rocks of the Front

Range of the Rocky Mountains clearly furnished most of the detritus of the Red Beds, and it is probable that the copper of these beds was derived from the weathering of the numerous primary copper ores found within the pre-Cambrian areas (14, p. 174). In the Nacimiento Mountains of New Mexico the cores of the ranges are made up of pre-Cambrian crystalline rocks which are known to contain copper ores of the more usual types although not in workable amounts (24, p. 149).

The Permian Red Beds of north-central Texas and of Oklahoma are composed in large part of debris derived from pre-Cambrian igneous rocks and earlier Paleozoic sediments of the Wichita and Arbuckle Mountains [8] whose uplift began near the middle and terminated near the close of Pennsylvanian times. The later red beds may be to a large degree the redeposited sediments of late Upper Pennsylvanian or early Permian age. Although no ore deposits of commercial importance have been found in the Wichita Mountains, pyrite, chalcopyrite, and less commonly other sulphides occur at scattered localities in amounts that have attracted the prospector. In the Arbuckle Mountains only insignificant mineralization has been noted. The pre-Cambrian rocks of the Llano-Burnet Region of Central Texas contain scattered copper deposits.[9] These rocks are not in direct contact with the Permian formations and are remote from recorded localities of copper in the Red Beds. Their contributions, if any, to the copper of the Red Beds must have been indirect through the erosion of earlier Paleozoic sediments.

Although some cupriferous debris from the older highlands must inevitably have been incorporated mechanically in the sediments flanking them, such contributions of the readily decomposed cupriferous minerals appear to have been of minor consequence and have not been recognized in any of the described

[8] Taff, J. A.: Preliminary Report on the Geology of the Arbuckle and Wichita Mountains in Indian Territory and Oklahoma, with an appendix on Reported Ore Deposits of the Wichita Mountains, by H. Foster Bain. U. S. Geol. Survey, Prof. Paper 31, 1904. Also Hoffman, M. G.: The Geology and Petrology of the Wichita Mountains. Doctor's Dissertation, University of Chicago, 1930.

[9] Paige, Sidney: Mineral Resources of the Llano-Burnet Region, Texas. U. S. Geol. Survey, Bull. 450, 1911, p. 73–74.

deposits. It is clear that the copper was transported almost wholly in solution and re-deposited in its present situations through the mechanism mainly of replacement and subordinately of the filling of openings.

Composition of the Transporting Solutions.—If the copper-bearing solutions responsible for the Red Beds type of deposit were derived from the weathering of older copper deposits of the neighboring highlands, what were the physical and chemical characteristics of these solutions? As these solutions formed part of the shallow ground-water circulation, their temperatures and pressures were moderate and the dissolving of copper and iron from the older copper deposits must have been accomplished in the ways that have already been outlined. Indeed, downward enrichment must have proceeded hand in hand with the contribution of copper and iron to the general ground-water circulation. The contribution of copper was presumably considerable. Some of the iron dissolved from the oxidized zone and most of that dissolved from the ground-water zone entered the general ground-water circulation. Both metals originally joined the ground-water circulation in balance mainly with the sulphate radicle. In contrast to the small lateral component of transportation in downward enrichment, in the formation of the Red Beds deposits, copper was transported for distances measured in scores or even in hundreds of miles, though probably not in a single step. It is desirable, therefore, to consider the chemical nature of the transporting ground water as an important factor in the problem of redeposition of the metals.

The dominant chemical characters or qualities of the ground waters were probably *not* determined by the relatively small contributions from the weathering of copper ores but rather by the mineral constitution of the sediments through which they circulated. As the composition of the sediments has undergone relatively little change from the time of their deposition to the present day, it is worthwhile to examine the nature of the existing ground waters. The present ground-water conditions are significant also in view of the possibility that in many of the deposits

the processes of ore deposition may still be operative. Variations in precipitation with changing climatic conditions are also to be reckoned with but they would be expected to affect the concentration more largely than the quality of the ground waters. The composition of the waters was such as to permit lengthy transport of the copper.

In the literature descriptive of the Red Beds deposits, few references to the composition of the waters in the sediments are to be found. Near the copper deposits of Arriba, New Mexico, are numerous salt springs, and at the Guadalupe deposits in the same State considerable water is reported at depths of 300 feet, unfit for use and containing gypsum. At Red Gulch, Colorado, the sparse waters are reported to contain abundant sulphates, and near the Cashin mine in the same State springs carry "salt or alkali." Of particular interest are the Springs at the former Imperial baths at Sergiefsk, Russia, in Permian beds some of which contain copper deposits. These springs are highly charged with hydrogen sulphide and analyses show that among the acid radicles the sulphate radicle is dominant with carbonate next, and among the basic radicles calcium is dominant and alkalies next. Some chloride radicle is also present.

In the literature dealing in a more general way with the Red Beds of the United States the quality of the underground waters is usually described in such general terms as saline, gypsiferous or alkali, indicating a mineral content commonly so high as to render the waters unsuitable for domestic or industrial use. The waters of the copper-bearing Permian rocks of north central Texas have fortunately been studied in some detail by C. H. Gordon,[10] and a number of analyses are available. The waters from wells in the Permian formations of this region are mostly highly mineralized, salty, and unfit for use. The composition of four of these well waters is represented graphically in Fig. 4, the analyses having been recalculated in terms of reacting values of the radicals, following the Palmer method.[11] A comparison of these graphs in-

[10] Gordon, C. H.: Geology and Underground Waters of the Wichita Region, North-central Texas. U. S. Geol. Survey, Water Supply Paper 317, 1913.

[11] Palmer, Chas.: The Geochemical Interpretation of Water Analyses. U. S. Geol. Survey, Bull. 479, 1911.

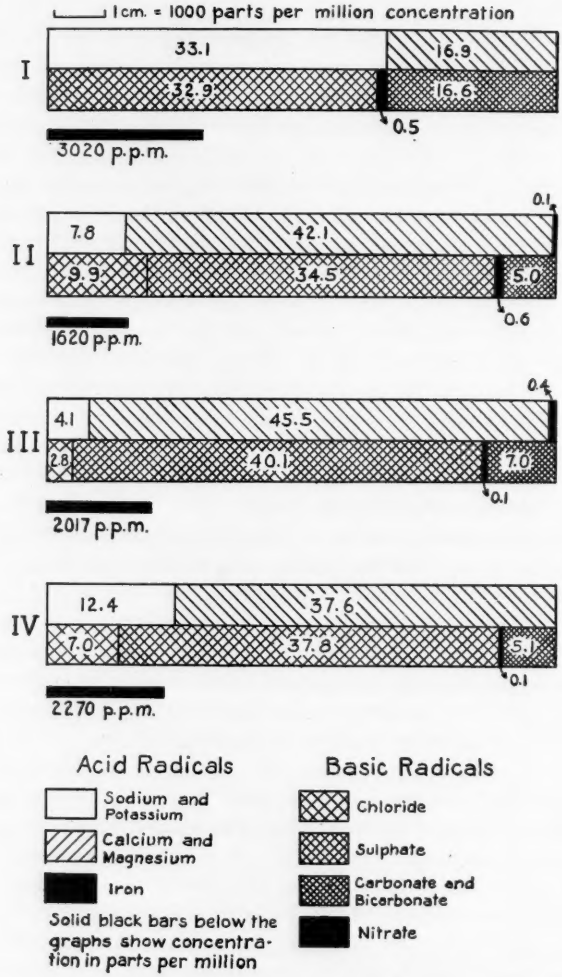

FIG. 4. Graphs showing qualities and concentrations of four ground waters from the Permian formations of North Texas.

dicates that the dominant basic radicles are alkalies and the alkali earths. In one of the samples alkalies are dominant, in the other three the earths are dominant. Among the acid radicles, the sulphate radicle is dominant in all the samples; the chlorine radicle is present in three of the samples. Weak acid radicles (carbonate and bicarbonate) are present in all the waters and are in part in balance with the earths, which feature should give to the waters an alkaline reaction. Among the metals, iron and aluminum are present in very small amounts. It should be noted that the four analyses are from three different counties and are probably representative of ground-water conditions over a considerable area.

Gould [12] has given a considerable number of analyses of well waters from the Permian Red Beds of Oklahoma. In these the alkalies were not separately determined and it is not practicable to present the results graphically. Among the acid radicles either chloride or sulphate may predominate. The carbonate radicle is commonly present but subordinate. Sodium chloride and gypsum may be so abundant as to render the waters unfit for use.

From these somewhat meager data it is not possible to generalize with safety, but the available data indicate that a high concentration of dissolved salts is common; that many of the waters have a high sulphate and some of them a high chloride content; weak acid radicles are present; and some waters at least are of alkaline reaction. In such waters it is to be expected that copper and iron would be transported dominantly in balance with the sulphate or chloride radicle.

The Deposition of the Iron.—The poverty of the deposits of the Red Beds type in iron as contrasted with the pyritic types of copper deposits implies its ultimate precipitation from the copper in different situations and presumably through different mechanisms. Iron and copper from the weathering of sulphide ore bodies enter the ground-water circulation, as has been pointed out, largely as ferrous sulphate and cupric sulphate.[13]

[12] Gould, C. H.: Geology and Water Resources of Oklahoma. U. S. Geol. Survey, Water Supply Paper 148, 1905, 143–148.

[13] This association is not a wholly stable one, for laboratory experimentation indicates that the equation $2FeSO_4 + 2CuSO_4 \rightleftarrows Fe_2(SO_4)_3 + Cu_2SO_4$ is in some

It is hardly to be expected that the long journey from the ancient land masses that were the sources of these metals to the places where they are found today was accomplished in a single step. It is to be pictured, probably, as many stages each involving in succession deposition, erosion, oxidation and solution, transportation and redeposition. If this concept is correct, sulphides of copper and of iron precipitated within the red beds were eventually exposed to oxidation. In the course of oxidation, as is well known, a part of the iron is likely to become fixed in relatively insoluble forms, such as " limonite," whereas copper is more largely dissolved or is fixed more transiently in oxidation products less stable than limonite. In so far as such selective processes are operative, the proportion of soluble copper salts as compared with soluble iron salts may increase as the long and complex process of transportation goes on.

If there is a tendency for iron to part company from copper by being precipitated in difficultly soluble forms, it may afford an explanation of those Red Beds chalcocite deposits which show no evidence that iron sulphides were ever among their components.

The Mechanism of Copper Deposition.—From considerations already set forth it seems probable that the copper reached its locus of deposition in solution in the ground waters mainly in balance with the sulphate and chloride radicles. As it was deposited mainly in the form of chalcocite, the process of deposition involved reduction, cupric sulphate ($CuSO_4$) being reduced to cupric sulphide (Cu_2S). What were the effective agents in this reduction?

In so far as nodules or concretions of pyrite, of marcasite, or of chalcopyrite were present in the sediments—whatever may have been their origin—such minerals were capable of serving as precipitating agents, reducing the cupric sulphate solutions and depositing copper through the mechanism of replacement in the

measure reversible. But at ordinary temperatures it appears that the quantities of ferric sulphate and of cuprous sulphate that can coexist with ferrous sulphate and cupric sulphate are small. See Zies, E. G., Allen, E. T. and Merwin, H. E.: Some Reactions Involved in Secondary Copper Sulphide Enrichment. ECON. GEOL., vol. XI, 1916, pp. 437–452.

form of chalcocite or more rarely of bornite, the chemical reactions involved being the familiar ones of downward enrichment. Microscopic studies of some of the ores show conclusively that this mechanism of deposition has been operative. Thus in the Blue Dike prospect in the White Canyon Region of Utah the oldest sulphides of the ores are pyrite and chalcopyrite and these have been replaced by covellite and chalcocite (19, p. 622). Near La Sal, Utah, the chalcocite is in part at least a replacement of pyrite and of chalcopyrite (19, p. 615). At Salt Wash, Utah, replacement remnants of chalcopyrite are abundant in some of the ores (19, p. 616). In the deposits of Cotton County, Oklahoma, the chalcocite is in part at least a replacement of iron sulphide—either pyrite or marcasite (10, p. 144). In the ores of the Sierra Oscura, Rogers has clearly shown replacements of pyrite by bornite and of the latter by chalcocite (3, p. 370). It is clear, therefore, that the types of reducing reactions through the agency of ferruginous sulphides, familiar in downward enrichment, have been responsible for a part of the deposition of chalcocite and bornite.

From most of the deposits of this type, however, ferruginous sulphides, other than bornite, have not been reported and chalcocite is the principal sulphide noted. Although it is likely that microscopic studies of polished specimens of more of these ores will reveal replacement remnants of ferruginous sulphides, it seems unsafe to conclude that none of the chalcocite and bornite have been directly precipitated from solution. On the contrary, it is a reasonable supposition that reducing agents that would precipitate iron sulphides from ferruginous sulphate solutions would be capable of precipitating chalcopyrite or bornite from solutions containing both iron and copper and of precipitating chalcocite from sulphate solutions relatively free from iron but rich in copper. The nature of the sulphides precipitated would be likely to depend largely, therefore, on the ratios of copper to iron in the solutions at the particular locality where precipitation was in progress.

The association of plant remains with many of the deposits of

this type is so intimate that it is extremely probable that they have exerted an influence in ore deposition. The possibility should be recognized that this influence may be more largely physical than chemical. The familiar phenomenon, sometimes associated with these deposits, of the selective replacement of wood by silica to form "petrified" wood is not a process for which any chemical equation can be written. The published descriptions do not record replacements of silicified wood by sulphides but if such occur they would have to be interpreted as physical substitutions. It is also possible that some of the observed direct replacements of carbonized plant tissues by sulphides are physical substitutions.

On the other hand, some or all of the replacements of plant tissues by sulphides may directly or indirectly involve chemical interaction and they have in fact commonly been so interpreted, carbon or some hydrocarbon having usually been postulated as the reducing agent which deprived sulphate solutions of the metals of their oxygen and led to the precipitation of the metal as a sulphide. It is well known that direct reductions of sulphates to sulphides through the agency of organic matter can take place at temperatures of 700°–1000° C. and certain commercial processes, such as the Le Blanc process of soda manufacture, depend upon such reactions. But it has been pointed out by the writer and by others [14] that thus far there is no experimental evidence that reductions of this sort can proceed at ordinary temperatures. Since it is quite possible that further experimentation will supply such evidence, it would be highly incautious at present to summarily dismiss the possibility of direct reduction.

On the other hand it is common knowledge that organic matter by functioning as a food for bacteria may lead through biochemical reactions to the production of hydrogen sulphide and probably other products of bacterial metabolism which are effective reducing agents. Hydrogen sulphide is a common product of the

[14] Allen, Crenshaw and Johnston: The Mineral Sulphides of Iron. Amer. Jour. Sci., Vol. XXXIII, 1912, p. 171. Bastin, E. S. et al.: The Problem of the Natural Reduction of Sulphates. Bull. Amer. Assoc. Petr. Geol., Vol. X, 1926, pp. 1281–1286. Thiel, Geo. A.: Experiments Bearing on the Biochemical Reduction of Sulphate Waters. ECON. GEOL., Vol. XXV, 1930, pp. 249–250.

putrefaction, through the agency of a variety of bacteria, of organic sulphur-bearing compounds in *animal* remains—as convincingly witnessed by the spoiling of eggs. It is commonly present as a minor component in the gases given off from sewage and would be more abundant in these gases were it not for its activity in combining with iron salts to form iron sulphide. Old sewage sludges are commonly dark in color from the abundance of iron sulphide.[15]

Fewer data are available concerning the development of hydrogen sulphide in the decay of *plant* tissues. Sulphur is a component of many proteins present in plants and, in the laboratory, plant extracts rich in such proteins have yielded hydrogen sulphide when inoculated with the same types of bacteria that yield this gas from animal extracts.[16] It would be reasonable to expect such processes to be operative in the early stages of plant decay when protein components are abundant. The odor of hydrogen sulphide is frequently noted where vegetation is decaying but it is usually uncertain whether the gas comes from the breaking down of organic sulphur compounds in the plant tissues or of inorganic sulphates dissolved in the waters in or around the plant remains. The fact that the odor of hydrogen sulphide is more commonly noted in salt marshes [17] than in fresh water marshes suggests the importance of the reduction of inorganic sulphates.

The writer and his associates have presented experimental evidence that the waters associated with oil even at considerable depths in certain oil fields in Illinois and in California contain living bacteria. The waters in which they occur are charged with hydrogen sulphide, commonly in abundance, and these bacteria when cultured in the laboratory function as sulphate reducers with the generation of hydrogen sulphide. Inorganic sulphates dissolved in the ground waters and the hydrocarbons of the oil are apparently necessary to their metabolism. It is a reasonable supposition supported by the abundance of sulphate-reducing bacteria in many muds rich in organic matter that *other* hydrocar-

[15] Fuller, Geo. W.: Sewage Disposal (1912), 92–95, 484–486.

[16] Stagnitta-Balistreri: Die Verbreitung der Schweifelwasserstoffbildung unter den Bacterien. Arch. für Hygiene, Bd. 16, 1893, 33–34.

[17] Grabau, A. W.: A Comprehensive Geology, Pt. I, p. 333.

bons such as those of plants may also in the presence of sulphate waters serve as food for such bacteria and result in the generation of hydrogen sulphide, one of the most effective precipitants of sulphides of the metals. In this connection it should be noted that Thiel [18] has obtained marked bacterial reduction of ferrous and zinc sulphates in cultures made from sediments rich in plant debris. David White [19] has pointed out that where wood has been replaced by sulphides the cells are not collapsed nor the stems flattened, indicating replacement prior to deep burial and therefore at a time when bacterial action would presumably be active.

Lovering [20] has proved that in thin beds of black muck in a bog near Cooke, Montana, native copper has been precipitated from copper sulphate solutions by the indirect effect of a variety of organic compounds resulting from bacterial metabolism; in this instance, however, H_2S was not one of the bacterial products and no copper was precipitated as sulphide.

Native copper is absent from most of the deposits of the Red Beds type and where present is in meager amounts only. It is probable that in such occurrences, as in the oxidized zones of so many chalcocite deposits, it is a residuum after the oxidation of the sulphur of chalcocite. The possibility that some may have been precipitated by biochemical reactions, such as those noted by Lovering must, however, be considered.

The Boleo copper deposit in Lower California (32) appears to be an interesting variant from the usual Red Beds type. Copper was deposited mainly as chalcocite by the replacement of clays interstratified with beds of volcanic tuffs. The copper is thought by Touwaide to have originally been in balance with the silicate radicle in the pyroxenes of the volcanic tuffs, to have been leached therefrom by connate waters, transported as sulphate or chloride, and reprecipitated as chalcocite through the agency of hydrogen sulphide or of some organic sulphur compound the origin of which is not discussed. None of the Boleo beds are red.

[18] Thiel, George A.: Experiments Bearing on the Biochemical Reduction of Sulphate Waters. ECON. GEOL., Vol. XXV, 1930, p. 247.

[19] White, David: Trans. Amer. Inst. Min. Met. Eng., Vol. 76, 1928, p. 386.

[20] Lovering, T. S.: Organic Precipitation of Metallic Copper. U. S. Geol. Survey Bull. 795, 1927, 45–52.

THE MANSFELD TYPE.

A third type of copper deposit somewhat *less* deficient in iron and sulphur as compared with the "standard" type is best represented by the copper shales of Mansfeld, Germany. These deposits, exploited commercially for seven and a half centuries, are the subject of a voluminous literature which has been summarized by Trask [21] and more recently and fully by Gregory.[22] The copper ore is a thin bed of Upper Permian shale that lies just above the Rotliegende (Lower Permian) and below the Zechstein—a bituminous limestone near the base of the Upper Permian. The ore bed is a bituminous, marly shale 5 to 20 inches thick and averaging 9 inches, extending over several thousand square miles. The mined ore carries 2.7 to 3.2 per cent. copper but the distribution of values is irregular and in parts of the Mansfeld district the copper content falls as low as 0.7 per cent., and large areas of the same shale in neighboring districts are barren of ore. The present bituminous content of the copper shale varies from 10 to 20 per cent.

The bulk of the copper occurs as *bornite;* chalcocite and chalcopyrite are also present but the microscopic evidence seems to indicate that they are commonly secondary. The copper minerals are mainly in minute particles irregularly disseminated or concentrated in lines or streaks parallel to the bedding, but nodular or concretionary forms constitute a minor mode of occurrence.

Among European geologists a spirited controversy has centered about the syngenetic versus the epigenetic character of these ores, but the trend of later opinion, influenced by microscopic studies of the ores and a growing appreciation of the potentiality of bacteria as agents of ore deposition, is strongly towards a dominantly syngenetic origin. The minor faults or "Rucken", that transect the copper shale and in and near which relatively rich ores may occur, are under the syngenetic interpretation regarded as chan-

[21] Trask, Parker D.: The Origin of the Ore of the Mansfeld Kupferschiefer, Germany: A Review of the Current Literature. ECON. GEOL., vol. XX, 1925, 746–761.

[22] Gregory, J. W.: The Copper Shale (Kupferschiefer) of Mansfeld. Inst. Min. and Met., Bull. 315, 1930, 28 pp.

nels along which some secondary concentration and redeposition has occurred and not as channels of primary mineralization. As stated by Gregory, "the influence of the "Rucken" is that of secondary enrichment," and many of the rich ore patches are not associated with "Rucken."

The underlying Rotliegende contains some igneous rocks, chiefly melaphyres and quartz porphyries of early Permian age, but they are older than the "Rucken" and they contain only minute amounts of copper. There is no evidence that they were the source of the copper either on the syngenetic or the epigenetic hypothesis.

It is particularly noteworthy that the copper shale is not a permeable horizon and that much more permeable Permian beds even where bituminous were *not* loci of ore deposition.

Bearing on the conditions of deposition of the copper shales, the fossil remains in the beds offer important evidence. The copper shale carries abundant leaves of land plants and reptilian and fish remains, but marine brachiopods, common in the overlying Zechstein, are absent. After a critical review of the fossil evidence Gregory concludes that "the Mansfeld Basin seems to have been a lagoon in which the fossils are land plants, land reptiles and fish which may have lived both in fresh and salt water." [23]

As to the genesis of the deposits, Gregory concludes that the original source of the copper was the weathering of older copper deposits in the Harz Mountains or other highlands flanking the "Kupferschiefer" lagoon, the copper being carried to the lagoon in the surface and ground-water circulation and there redeposited. Gregory suggests that it was redeposited first in the form of carbonates which were later converted to bornite "when the shale was saturated with some sulphurous acid, formed either from sulphur obtained from reduction of sulphate in the water or from the organic matter in the deposits on the floor of the lagoon." It seems equally as likely that the metals were deposited directly as sulphides through the agency of hydrogen sulphide generated by

[23] *Ibid.*, p. 20.

the bacterial reduction of sulphates in the presence of organic matter, in a fashion analogous to the well established deposition of pyrite in the muds of the Dutch canals and elsewhere.

The Mansfeld type presents analogies to the Red Beds type in the probable derivation of its copper from the weathering of older deposits of the " standard " type and its transport by ground waters and in the probable rôle of hydrogen sulphide in its redeposition, but contrasts with the latter in the deposition of the copper in lagoonal rather than terrestrial sediments and in a lesser degree of separation of copper from iron.

THE LAKE SUPERIOR TYPE.

In contrast to the copper deposits previously described, are those of which Lake Superior and probably also Corocoro, Bolivia, are examples. They are characterized by native copper as their dominant ore mineral and are sufficiently closely associated with igneous rocks to suggest a magmatic source for the mineralizing solutions. They are, however, mostly associated with red rocks and this association has led to some confusion with the meteoric Red Beds type of deposit.

In the Lake Superior deposits the persistence in depth with unchanged character probably precludes a surficial origin for the mineralizing solutions. Moreover the deeper levels of all the mines are essentially dry and the mineralogy of the ores is such as to indicate moderately high temperature of ore formation.

The Lake Superior deposits are fillings and replacements in conglomerates and in vesicular basalts. Although native copper is the dominant cupriferous mineral, copper sulphides and arsenides are widespread and in order of abundance include chalcocite, bornite and chalcopyrite. It is noteworthy that pyrite is practically absent. The arsenides present are those richest in copper, and some metallic arsenic occurs in alloy with copper. Arsenic is much more abundant in proportion to sulphur than in other copper districts.[24]

[24] Butler, B. S., Burbank, W. S., et al.: The Copper Deposits of Michigan. U. S. Geol. Survey, Prof. Paper 144, 1929.

The association of mineralization with basic volcanic rocks is intimate. Most of the ore occurs in vesicular flows of basalt, and the central portions of some of the thickest of the flows contain pegmatitic knots and lenses carrying copper sulphides and native copper. The exposed plutonic rocks of the region (notably the gabbros of Mt. Bohemia) carry at numerous points veins of chalcopyrite, chalcocite, bornite, and rare pyrite and pyrrhotite. The Duluth gabbro, believed to be of late Keweenawan age, carries numerous deposits of copper and iron sulphides. Small amounts of chalcopyrite and pyrrhotite occur in this gabbro at many points and also in the contact metamorphic zones bordering it.[25] There is reason to believe that the Duluth gabbro extends beneath the copper district, and it is regarded by Butler and his associates as the probable source of the solutions that deposited the native copper ores.

At Corocoro, in Bolivia,[26] as at Lake Superior, native copper is the dominant cupriferous mineral although chalcocite is relatively more abundant. Pyrite and chalcopyrite are absent. The environmental conditions are, however, notably different from those at Lake Superior. The ore forms fillings and replacements mainly in sandstones which are interbedded with shales, conglomerates, marls and some beds of gypsum and rock salt. Many of the beds are red and some contain plant remains. Thus, there are some resemblances between the environment at Corocoro and that of some of the Red Beds type. There appears to be no evidence, however, that the plant remains have played any consequential rôle in the mineralization. The ore deposition was accompanied by bleaching of the red sediments—a feature not a characteristic of the Red Beds type.

Igneous rocks are absent from the immediate vicinity of the Corocoro mines but they occur 15 km. to the west and 20 km. to the north and include both acid and basic types. There are, however, chemical considerations pointing to a magmatic source for the ore-depositing solutions.

[25] Nebel, M. L.: The Basal Phases of the Duluth Gabbro near Gabamichigami Lake, Minnesota, and its Contact Effects. ECON. GEOL., vol. XIV, 1919, p. 399.

[26] Singewald, J. T., Jr., and Berry, E. W.: The Geology of the Corocoro Copper District of Bolivia. Johns Hopkins University Studies, No. 1, 1922.

Assuming tentatively a magmatic source for the mineralizing solutions at Lake Superior and at Corocoro, the question arises, were the solutions poor in iron and sulphur when they left the magma or were these elements subsequently removed from the solutions? To answer this fundamental question requires (1) evidence of the composition of the mineralizing solutions at the time they left the probable magmatic source, and (2) evidence of interaction between wall rocks and mineralizing solutions that might modify the composition of the latter.

Evidence of the first is afforded by the mineralization observed in the Duluth gabbro and its contact zones, and in the probably contemporaneous gabbro of Mt. Bohemia. The copper mineralization associated with these is all of the "standard" type, copper being present as chalcopyrite in association with pyrite, pyrrhotite, or both. If the gabbro was the source of the mineralizing solutions, there was apparently no deficiency in iron and in sulphur at the time the solutions left the magmas and entered the wall-rocks of the immediate contact zones.

Before passing to the second question, it is appropriate to enquire as to the nature of the changes in the rocks that might be competent to produce the results observed. Iron and sulphur, if present in their usual abundance in the mineralizing solutions as they left the magma, must have been removed as readily soluble compounds, presumably as sulphates. The formation of iron sulphates from iron in balance with the sulphur radicle is a process of oxidation necessitating the relinquishment of oxygen by some component of the rocks penetrated by the mineralizing solutions, *i.e.*, the reduction of some mineral or minerals of the rock.

Of the *commoner* rock-forming elements Si, Al, Na, K, Ca, Mg and Fe, all except iron and silicon have only one valence and therefore are capable of combining with oxygen in only one proportion. Although silicon has more than one valence it appears to be tetravalent in all of the common rock-forming minerals. Thus when the available oxygen, as determined by analysis, is allotted, in the proportions required by their known valences, to

the other elements present, the remaining oxygen is sufficient to combine with silicon as SiO_2. Iron appears to be the only one of this group of elements which is characterized by more than one valence in its occurrences in the common rocks, and hence appears to be the only one capable of playing an important rôle in oxidation phenomena in the native copper deposits. Thus the state of oxidation of the iron-bearing minerals of the rocks associated with the native copper ores must be investigated to see if there is available oxygen to combine with iron and sulphur in the mineralizing solutions.

Ferric iron is abundantly present as oxides in the rocks associated with the native copper deposits of Lake Superior and Corocoro, as evidenced by the prevailing red color of many of the beds, and attention has naturally been focussed upon ferric oxide as a potential oxidizing agent. Although it may well have been the principal agent concerned, it seems possible that it was not the sole agent in the Lake Superior district, and that in some other districts its rôle may have been subordinate. At all events it should not be lost sight of that iron is abundantly present in *silicates* in association with many occurrences of native copper, and that in these silicates it may be present in either the ferrous or ferric form.

A survey of the available analyses as tabulated by Hintze [27] and by Doelter [28] shows that in most of the silicates (with the exception of biotite), common in unaltered igneous rocks, iron is present either wholly or dominantly in the ferrous form. Thus in the olivines and in hypersthene the iron is essentially wholly ferrous. In augite and hornblende ferrous iron is usually dominant. In biotite no consistent relationship is discernible, but biotite is only a rare component of the basic volcanic rocks associated with native copper deposits.

Among the *secondary* iron-bearing minerals common in igneous rocks, particularly of the basic types, most of the chlorites show a great preponderance of ferrous iron. In the mineral epidote, on the other hand, nearly all of the iron is ferric. Thus in 54

[27] Hintze, C.: Handbook der Mineralogie, Bd. II.

[28] Doelter, C.: Handbook der Mineralchemie, Bd. II, 1914.

analyses of epidotes cited by Doelter the ferrous iron is mostly less than 1 per cent., whereas ferric iron ranges from 1.68 to 20.24 per cent., and ferrous iron exceeds ferric in only four analyses. In the 61 analyses tabulated by Hintze ferric iron is greatly in excess in all but two. This dominantly ferric state of the iron is recognized in the generalized formula $4CaO.3(Al.Fe)_2O_3.6SiO_2.H_2O$.

That iron may change its state of oxidation not only when combined in oxides but also when combined in silicates is clearly shown by the phenomenon of epidotization as illustrated by the following analyses of a felsite porphyry and of the epidosite derived from its alteration.[29]

FELSITE PORPHYRY (I) AND EPIDOSITE (II) DERIVED FROM ITS ALTERATION. ISLAND OF HOCHLAND.

	I	II
SiO_2	74.46	71.98
Al_2O_3	12.55	12.55
Fe_2O_3	0.78	3.43
FeO	2.95	0.83
MgO	0.20	0.68
CaO	1.83	9.81
Na_2O	0.54	0.46
K_2O	6.69	0.26
	100.00	100.00

Is it possible that under other circumstances the highly oxidized silicate epidote may be reduced and thus function as an oxidizing agent in the formation of some native copper deposits? Epidote is one of the abundant minerals of the Lake Superior deposits. It is particularly interesting to note that Watson [30] in discussing the origin of the economically unimportant native copper associations with basaltic lavas in the South Atlantic States, pointed out that red phases of the basalts were absent and that it was therefore unlikely that ferric oxide was the agent of oxidation. To quote his description: "The most striking feature in the composition

[29] Lemberg, J., in Doelters Handbook, Bd. II, p. 854.

[30] Watson, T. L.: Native Copper Deposits of the South Atlantic States Compared with those of Michigan. ECON. GEOL., vol. XVIII, 1923, pp. 737–739.

of the lavas at present is the extensive development of chlorite and epidote through chemical metamorphism of the principal minerals, pyroxene and plagioclose," and he concluded that " it seems most probable that there is a genetic relationship of the copper to epidotization, since the native metal is usually found in the more highly epidotized portions of the lavas." Another association of native copper with pronounced epidotization is in the Comobabi Mts. of Arizona.[31]

If, as has been postulated, iron, copper, and sulphur were present in the original mineralizing solutions before their modification in the proportions necessary to form chalcopyrite ($CuFeS_2$ or $Cu_2S.Fe_2S_3$) and pyrite (FeS_2), and if, through interaction with the country rocks, sulphur and iron were removed as ferrous sulphate (Fe_2SO_4), it follows that not oxygen alone but also iron was contributed by these rocks, since the ratio of iron to sulphur in ferrous sulphate is greater than in pyrite and chalcopyrite. The same holds true in lesser degree for ferric sulphate. In view of the abundance of calcite and the rarity of gypsum in the ores, it is unlikely that much sulphur was removed as sulphuric acid.

With these fundamental considerations in mind we may next examine the field evidence of alteration of the country rocks in the Lake Superior region and at Corocoro. In the Lake Superior region Butler and associates have shown that hematite is abundant in the conglomerates and in the upper parts of the basaltic flows and antedates the mineralization. The red color of these rocks is due to its presence. Where copper is abundant, conspicuous bleaching of these rocks has occurred, and analyses of bleached and unbleached phases show a pronounced decrease in ferric iron in the bleached portions and also a decrease in total iron. These relations, as has been clearly pointed out by Butler and his associates, are harmonious with the concepts just outlined and show that ferric oxide of the rocks has not only served as an oxidizing agent but that a large part of the iron involved in the reaction was removed in solution.

[31] Butler, B. S., Burbank, W. S., et al.: *Op. cit.*, p. 144.

Although it is probable that ferric oxide was the dominant source of oxygen and of iron in the reactions resulting in the deposition of the Lake Superior ores, the possible rôle of epidote as a subsidiary oxidizing agent in that region and possibly the dominant agent in certain other districts requires consideration. To quote Butler and his associates,[32] epidote " is everywhere abundant in both veins and lodes, being perhaps especially abundant in the Evergreen and succeeding lodes of that series and in the Isle Royale lode. It is also abundant locally in the Calumet and Hecla conglomerate. Epidote characteristically has replaced the rock minerals, though it has also been formed in openings. . . . Epidote is one of the early minerals of the ore-forming period, in the main preceding copper, but it also appears to have been formed to a slight extent later than copper."

If epidote served as an additional source of iron and oxygen, silicates somewhat analogous to epidote in composition, but relatively poor or lacking in iron, that might be products of its alteration, should be present in the ores. A number of such minerals are present, and the conventionalized formulae of those that are abundant may be listed in comparison with epidote:

Epidote ..... $4CaO.3(AlFe)_2O_3.6SiO_2.H_2O$
Pumpellyite ..... $6CaO.3Al_2O_3.7SiO_2.4H_2O$
Prehnite ..... $2CaO.Al_2O_3.3SiO_2.H_2O$
Laumontite ..... $CaO.Al_2O_3.4SiO_2.4H_2O$

Pumpellyite is the mineral that gives the characteristic bluish-green color to much of the lode material in the mines in the south end of the district. To quote Butler:[33] " The mineral is abundant in all the amygdaloid lodes and in all the fissures. It is present but not usually abundant in the conglomerates. . . . It replaced the earlier minerals of the ore period such as chlorite and epidote, and in turn was replaced by later minerals, including copper." According to Palache[34] some of the pumpellyite was a

[32] *Ibid.*, p. 60.

[33] *Ibid.*, p. 60.

[34] Palache, Chas. and Vassar, Helen E.: Some Minerals of the Keweenawan Copper Deposits. Amer. Miner., vol. X, 1925, pp. 412–415.

late magmatic mineral antedating the copper mineralization. While iron does not appear in the formula of the mineral given by Palache and Vassar, the analyses show that iron oxides are present to the extent of between 7 and 8 per cent.

Prehnite, to quote Butler,[35] " is abundant in veins throughout the district. It is present in most of the amygdaloids, where it has filled vesicles or replaced earlier minerals or rock. . . . It is rare in the conglomerates. . . . In the veins it is very commonly associated with copper. . . . The highly prehnitized amygdaloid, however, is usually not high in copper."

It seems possible that the formation of pumpellyite, prehnite, and perhaps other minerals at a later stage than that at which epidote was deposited, is symptomatic of a gradual impoverishment of the deposits in iron and that certain components of the replaced epidote entered into their formation, but detailed microscopic study of their textural relations to epidote in the ores is essential to the validation of this concept, which is recorded here simply as a working hypothesis.

Watson's studies of the native copper ores of the South Atlantic States have already been mentioned but his conclusions with respect to their genesis may now be summarized in his own words:[36]

> Free ferric oxide in the form of hematite, an important constituent in the lava tops in the Michigan copper region, occurs only sparingly in the Appalachian lavas. . . . Metallic copper, the most important ore mineral of the region, was probably precipitated from sulphide solutions which entered the more highly epidotized portions of the rock along sheared zones rich in ferric iron as epidote. Ferric iron oxidized the sulphur in the solutions and deposited copper as native metal.

In Corocoro, Bolivia, the prevalent red color of many of the beds of the copper-bearing series indicates the abundance of hematite. Nearly all observers are in agreement in emphasizing the conspicuous bleaching of the beds where they are mineralized. To quote Singewald and Berry,[37]

[35] *Op. cit.*, p. 61.
[36] *Op. cit.*, p. 746.
[37] *Op. cit.*, p. 64.

Wherever mineralized, the beds have lost their red color and are white or light green. Residual patches of red rock are unmineralized and the impervious shales which enclose the cupriferous sandstones retain their red color. The bleached rock is as a rule soft and friable as though much of the cement that held together the sand grains had been removed.

The field evidence is strong that at Corocoro, as at Lake Superior, ferric minerals in the country rocks reacted with the mineralizing solutions to produce copper ores notably poor in iron and sulphur. *In the deposits of this class that occur in red sediments the association of the mineralization with bleaching of the rocks becomes a criterion differentiating them from copper deposits of the Red Beds type, in which bleaching if it occurs at all is a consequence not of mineralization but of the presence of organic matter.*

THE KENNECOTT TYPE OF COPPER DEPOSIT.

The last type of copper deposit characterized by rich copper minerals and poor in iron and sulphur that requires our consideration is represented by the great chalcocite deposits in limestone at the Kennecott Mines in Alaska. Although the geology of this and neighboring deposits has been searchingly studied by Bateman [38] and McLaughlin [39] and by Moffit,[40] its mode of origin is still not clear.

The oldest rocks of the region are interbedded sediments and basic lava flows of the Strelna formation—probably of Mississippian age. Unconformably overlying these is a thickness of not less than 4,000 feet of basic lava flows known as the Nikolai greenstone (Permian or Triassic), and above these without apparent unconformity lies the Chitistone limestone of Upper Triassic age.

Small copper deposits occur in the basaltic flows of the Strelna

[38] Bateman, A. M.: A Kennecott Type of Copper Deposit, Glacier Creek, Alaska. Econ. Geol., vol. XXVII, 1932, pp. 297–306.

[39] Bateman, A. M. and McLaughlin, D. H.: Geology of the Ore Deposits of Kennecott, Alaska. Econ. Geol., vol. XV, 1920, pp. 1–80.

[40] Moffit, Fred H.: The Metalliferous Deposits of Chitina Valley. U. S. Geol. Survey, Bull. 755, 1924, pp. 57–72.

formation and in the Nikolai greenstone but the ores of great economic importance are in the Chitistone limestone.

The copper deposits in the basic volcanic rocks of the Strelna and Nikolai formations occur exceptionally as contact metamorphic deposits adjacent to the granodiorites but mainly as replacements and fracture fillings remote from such contacts. These ores are small low-grade copper deposits commonly characterized by pyrite and chalcopyrite but some consist largely of bornite and chalcopyrite, still others of bornite and chalcocite or of chalcocite, and there are also many occurrences of native copper chiefly as amygdules. While some of the chalcocite and native copper are undoubtedly alteration products of earlier copper minerals, the microscopic evidence indicates that much may be primary. In the fracture fillings quartz, calcite and epidote are the principal gangue minerals and in the replacement deposits the basic volcanics are usually much altered with the development of chlorite, epidote, albite, calcite, and zeolites. Thus in the basic volcanic rocks a trend is indicated towards the development of copper ores relatively poor in iron and sulphur.

The great ore bodies of the district, those of the Kennecott Mines, are chalcocite ores in the Chitistone limestone. These ores are replacement veins and irregular replacement bodies related to fractures produced by folding. The roots of some of the deposits penetrate slightly into the underlying Nikolai greenstone.

The granodiorite intrusives are the only igneous rocks now exposed in the region that could have been the source of the mineralizing solutions. The presence of a granodiorite dike cutting the ore at the Kennecott Mine shows that some granodiorite formed later than the ores but it does not preclude an earlier origin for other bodies of granodiorite. The presence in the Kennecott ores of minor amounts of such minerals as enargite, tennantite and luzonite indicates moderately high temperatures of mineralization and probably implies an igneous source.

If the Kennecott ores were deposited from hydrothermal solutions depleted in iron and sulphur during their journey from a magmatic source to their place of deposition in the limestone, the

Nikolai greenstones are the most probable agent of such modification.

Obviously we are not yet in a position to discuss with assurance the genesis of the extraordinary deposits of Kennecott. It may be profitable to point out, however, that they contrast with the types of copper deposits previously considered (1) in their deposition in limestone, (2) in the absence of native copper, and (3), as has been emphasized by Lasky,[41] in the presence in some of the ore of textures indicative of deposition from colloidal solutions.

Deposition from colloidal solutions imposes certain restrictions as to genesis as compared with the usual ore types and removes other restrictions. First, it implies the presence in the mineralizing solutions of some dispersing agent permitting the products of chemical interaction to remain in fine mechanical suspension for long periods and to be transported in such suspensions possibly for considerable distances. This dispersing agent may itself have entered only partially, if at all, into the composition of the ore, making its nature a matter of inference or of speculation. Secondly, the difference in susceptibility of various components of a mineral-bearing solution to colloidal dispersion—and this difference in susceptibility may be large—introduces the possibility of notable changes in composition during the transport of the solution through the rocks. Thirdly, new mechanisms of precipitation may become operative, notably flocculation through the influence of electrolytes introduced from rocks with which the mineralizing solutions come in contact or through mingling with other solutions.

SUMMARY AND CONCLUSIONS.

The types of copper deposits that are most clearly and closely related to vulcanism, namely the magmatic type, the contact-metamorphic type, and the commoner hydrothermal types, although differing among themselves, are all characterized by an abundance of iron and of sulphur as evidenced by the prevalence

[41] Lasky, S. G.: A Colloidal Origin of Some of the Kennecott Ore Minerals. ECON. GEOL., vol. XXV, 1930, pp. 737–757.

of chalcopyrite and of pyrite or pyrrhotite. Their mineral composition is thought to be indicative of an abundance of iron and sulphur in the solutions from which they were deposited, as given off by magmas, whether these magmas were acid, intermediate or basic. Such copper deposits may be regarded as a "standard" type useful for comparison with other types.

The several types of copper deposits that form the subject of this paper are characterized by a dominance of the *richer* copper minerals, notably native copper, chalcocite and bornite, and by the absence or extreme rarity of chalcopyrite and pyrite. These types include (1) chalcocite zones of downward enrichment, (2) copper deposits of the "Red Beds" type, (3) the Mansfeld type of copper deposit, (4) the Lake Superior and Corocoro type, and (5) the Kennecott type. The first three are believed to be the result of the removal of iron and sulphur from *ores* of the "standard" type through gradational processes operating at or near the surface. The Lake Superior and Corocoro types appear to have resulted from the removal of iron and sulphur from copper-bearing *solutions* from magmatic sources through interaction with wall-rocks. The Kennecott type may belong here also, but conclusive data are as yet lacking.

The chalcocite zones of downward enrichment are well known to result from the oxidation of the "standard" type of copper ores, copper, iron and sulphur going into solution mainly as sulphates. Impoverishment in iron and sulphur is due predominantly to selective precipitation. Iron is in the main not reprecipitated within the ore body but is carried away as ferrous sulphate into the general ground-water circulation. Much of the copper on the contrary is reprecipitated, as chalcocite, or covellite, mainly through the replacement of primary sulphides within the ground-water zone. Native copper is not abundant in these deposits and where present is commonly a residuum from the oxidation of the sulphur of chalcocite.

The copper of the Red Beds type of deposit appears likewise to have been derived from the weathering of copper deposits of the "standard" type, occurring in older highlands which were

flanked by sediments mainly of terrestrial origin made up of the disintegration debris of these highlands. The metals copper and iron went into solution dominantly as sulphates. Mechanical contributions of copper minerals to the sediments seem to have been of minor consequence. The copper of the Red Beds represents that modicum of the dissolved copper that was *not* reprecipitated within the ore body but, along with iron, entered the general ground-water circulation and thus became available for transport to considerable distances.

The agency of transportation was ground water which appears to have been of the saline types characteristic of arid or semi-arid conditions, the waters being rich in sulphates and, in some instances, in chlorides. The initial balance of the metals with the sulphate radicle was therefore probably maintained during transport and it is possible that the sulphate and chloride salinity of the ground waters was an important factor in facilitating transportation.

Iron is conceived to have been more readily removed from solution than copper because of the facility with which ferrous sulphate oxidizes to ferric sulphate and finally to limonite, wherever such solutions come in contact with free oxygen. Iron was probably reprecipitated also as siderite. That some of the iron escaped precipitation in these oxidized forms is shown by the occasional presence of pyrite, marcasite or chalcopyrite in these deposits.

The precipitation of copper and iron as sulphides from sulphate or chloride solutions requires the action of some *reducing* agent or agents. None of the ordinary components of the ground waters or of the country rocks seems capable of functioning in this capacity at the ordinary temperatures involved, and appeal must apparently be made to the direct or indirect action of the organic remains—mainly of plants—commonly associated with these deposits. Although carbon and hydrocarbons can function as *direct* reducing agents at high temperatures, laboratory experimentation to date has failed to show that they can function similarly at ordinary temperatures. It would be injudicious at present, however, to rule out that possibility.

However, the organic matter may have acted as food for hydrogen sulphide-generating bacteria, which in turn yielded hydrogen sulphide that acted as a precipitating agent for the copper sulphide. Hydrogen sulphide is well known to be one of the products of putrefaction bacteria and it is also well established that, in the presence of organic food materials and of sulphate waters, sulphate reducing bacteria may be active in the formation of hydrogen sulphide. This gas is believed to have been the principal reducing agent operative in the precipitation of sulphides in the Red Beds deposits. Most of the chalcocite appears to have been a direct precipitation from solution through this agency but there is valid microscopic evidence that in some instances pyrite, marcasite or chalcopyrite were first precipitated and that these later served as precipitating agents for chalcocite or bornite after the fashion familiar in downward enrichment. It should be emphasized that hydrogen sulphide is a mobile reagent which may function at some distance from the organic matter about which it originated, and in this fact lies perhaps the explanation of chalcocite occurrences unassociated with plant remains.

The copper deposits of the Mansfeld type are interpreted as deposits in the bottom muds of shallow but extensive lagoons into which ground waters of much the same composition as those in the Red Beds have debouched. The agent of deposition was probably also hydrogen sulphide derived mainly from the bacterial reduction of sulphates in the presence of organic food material. The copper minerals later underwent some rearrangement through processes analogous to those of downward enrichment.

The three processes of (1) downward enrichment, (2) deposition from ground waters within bodies of sedimentary rocks, and (3) deposition from the waters of shallow lagoons, may be looked upon as successive and related stages in the ultimate redeposition of the products of weathering of the " standard " types of copper deposits. Deposition in each case involved the reduction of highly oxygenated copper solutions. Deposits of these three origins may be looked upon as forming one family group among the deposits characterized by rich copper minerals.

The copper deposits of the Lake Superior and Corocoro types contrast markedly with the above in the absence or rarity of organic remains, in the relative abundance of native copper, in the presence of accessory minerals indicative of high temperatures, and in the evidence of bleaching through the interaction of ferric oxide in the rocks with the mineralizing solutions. The available evidence favors the view that the mineralizing solutions came from magmatic sources and were already in a reduced condition, the metals being mainly in balance with sulphur. It is likely that they carried both iron and sulphur in abundance. From these initial solutions essentially all of the iron and much of the sulphur was removed through the oxidizing action of ferric oxide, and went into solution as iron sulphates. It is possible that in some deposits ferric silicates may have functioned as oxidizing agents with or instead of ferric oxide. The mechanism of ore deposition therefore contrasted markedly with that characteristic of the types of deposits previously described.

The ores of Kennecott, Alaska, appear to have been deposited from hot solutions presumably of magmatic derivation, although the nature and position of the igneous source is unknown. The process of oxidation of the solutions seems to have been in progress within the basic volcanics and may have reached its culmination in the chalcocite deposits in the limestone. The agents of oxidation were possibly ferric minerals in the basic volcanics, but specific evidence is lacking. The presence of colloform textures in some of the Kennecott ores probably indicates deposition from colloidal suspensions and the operation of some factors not involved in the genesis of the other ore types we have considered.

Copper ores of the Lake Superior and Corocoro type and possibly the Kennecott type constitute a second family group among the ores carrying rich copper minerals and deficient in iron and sulphur.

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BIBLIOGRAPHY OF THE MORE IMPORTANT PAPERS ON COPPER DEPOSITS OF THE RED BEDS TYPE.

General.

1. Stelzner, A. W., Bergeat, A.: Die Erzlagerstätten, vol. I, 1904, 418–442.
2. De Launay, L.: Gites Mineraux et Metalliferes, vol. II, 1913.
3. Rogers, A. F.: Origin of Copper Ores of the "Red Beds" Type. ECON. GEOL., vol. XI, 1916, 366–380.
4. Finch, John W.: Sedimentary Metalliferous Deposits in the Red Beds. Trans. Amer. Inst. Min. Met. Eng., vol. LXXVI, 1928, 378–392.

Texas and Oklahoma.

5. Schmitz, E. J.: Copper-Ores on the Permian of Texas. Trans. Amer. Inst. Min. Eng., vol. XXVI, 1896, 97–108.
6. Louis, Henry: Copper-Ores in the Permian of Texas. Discussion of the Paper by E. J. Schmitz. Trans. Amer. Inst. Min. Eng., vol. XXVI, 1896, 1051–1052.
7. Haworth, E. and Bennett, J.: Native Copper near Enid, Oklahoma. Bull. Geol. Soc. Amer., vol. XII, 1900, 2–4.
8. Tarr, W. A.: Copper in the "Red Beds" of Oklahoma. ECON. GEOL., vol. V, 1910, 221–226.
9. Richard, Louis M.: Copper Deposits in the "Red Beds" of Texas. ECON. GEOL., vol. X, 1915, 634–650.
10. Fath, A. E.: Copper Deposits of the "Red Beds" of Southwestern Oklahoma. ECON. GEOL., vol. X, 1915, 140–150.
11. Reiter, A. F.: Present Status of Copper Mining in Garfield County, Oklahoma. Oklahoma, Univ. Bull. (Proc. Oklahoma Acad. Sci. 1910–1920.) New Ser. 220, 1921, 67.

Colorado.

12. Emmons, S. F.: Copper in the Red Beds of the Colorado Plateau Region. U. S. Geol. Survey, Bull. 260, 1905, 221–232.
13. Emmons, W. H.: The Cashin Mine, Montrose County, Colorado. U. S. Geol. Survey, Bull. 285, 1906, 125–128.
14. Lindgren, W.: Notes on Copper Deposits in Chaffee, Fremont, and Jefferson Counties, Colorado. U. S. Geol. Survey, Bull. 340, 1908, 170–174.

Utah.

15. Rolker, C. M.: The Silver Sandstone District of Utah. Trans. Amer. Inst. Min. Eng., vol. IX, 1881, 21–33.
16. Boutwell, J. M.: Vanadium and Uranium in Southeastern Utah. U. S. Geol. Survey, Bull. 260, 1905, 200–210.

17. Hill, J. M.: Notes on the Northern La Sal Mts., Grand County, Utah. U. S. Geol. Survey, Bull. 530, 1913, 118.
18. Hess, F. L.: Carnotite near Green River, Utah. U. S. Geol. Survey, Bull. 530, 1913, 161–164.
19. Butler, B. S., Loughlin, G. F., Heikes, V. C. and Others: The Ore Deposits of Utah. U. S. Geol. Survey, Prof. Paper 111, 1920.

New Mexico.

20. Newberry, J. S.: Geological Report: Exploring Expedition from Santa Fe, New Mexico, to the Junction of the Grand and Green Rivers . . . in 1859, 1876. 67–69.
21. Peters, E. D.: Notes on the Oscura Copper-Fields and Other Mines in New Mexico. Eng. and Min. Jour., vol. XXXIV, 1882, 270–272.
22. Herrick, C. L.: The Occurrence of Copper and Lead in the San Andreas and Caballo Mountains (New Mexico). Amer. Geol., vol. XXII, 1898, 285–291.
23. Turner, H. W.: The Copper Deposits of the Sierra Oscura, New Mexico. Trans. Amer. Inst. Min. Eng., vol. XXXIII, 1903, 678–681.
24. Lindgren, W., Graton, L. C. and Gordon, Chas. H.: The Ore Deposits of New Mexico. U. S. Geol. Survey, Prof. Paper 68, 1910.
25. Gregory, Herbert E.: Geology of the Navajo Country. U. S. Geol. Survey, Prof. Paper 93, 1917, 140–141.
26. Stauber, I. J.: A Sandstone Copper Deposit (New Mexico). Mining Cong. Jour., vol. XVI, 1930, 929–931.

Arizona.

27. Hill, J. M.: Copper Deposits of the White Mesa District, Arizona. U. S. Geol. Survey, Bull. 540, 1914, 159–163.

Others, North America.

28. Weed, W. H.: Copper Deposits of New Jersey. New Jersey Geol. Survey, Ann. Rept. for 1902 (1903), 125–139.
29. Lewis, J. Volney: Copper Deposits of the New Jersey Triassic. Econ. Geol., vol. II, 1907, 242–257.
30. Wherry, E. T.: The Newark Copper Deposits of Southeastern Pennsylvania. Econ. Geol., vol. III, 1908, 737–738.
31. Gale, Hoyt, S.: Geology of the Copper Deposits near Montpelier, Bear Lake County, Idaho. U. S. Geol. Survey, Bull. 430, 1910, 112–121.
32. Touwaide, Marcel E.: Origin of the Boleo Copper Deposit, Lower California, Mexico. Econ. Geol., vol. XXV, 1930, 113–144.

33. Papenfus, E. B.: "Red-Bed" Copper Deposits in Nova Scotia and New Brunswick. Econ. Geol., vol. XXVI, 1931, 314–331.

Europe.

34. Murchison, R. I., de Verneuil, E. and von Keyserling, A.: The Geology of Russia in Europe, vol. I, 1845.
35. Neubert, Eduard: Die Kupfererzlager der Kargalinskischen Steppe im russischen Gouvernement Orenburg. Berg und Huttenmaennische Zeitung, Jahrg. XXII, 1863, 141–145.
36. Simon, C.: Kupfer-und Bleiablagerungen im bunten Sandsteine und Vogelsandstein der Umgegend von Saarlouis und St. Avold. Berg und Huttenmaennische Zeitung, Jahrg. XXV, 1866, 412–415.
37. Förster, C. R.: Notizen über den Kupferbergbau der Karlinski'schen Steppe im russischen Gouvernement Orenburg. Berg-und Huttenmaennische Zeitung, Jahrg. XXVII, 1868, 194–195.
38. Reh, Adolphe: Das Kupfererz und Salz Vorkommen in der Permischen Formation Südrussland. Zeitsch. für das Berg-, Hütten- und Salinen-Wesen, Bd. 29, 1881, 276–280.
39. Hering, C. A.: Die Kupfererzlagerstätte der Dyas in nordöstlichen Böhmen in Bezug auf ihre Abbauwürdigkeit. Oesterreichische Zeitsch. für Berg-und Huttenwesen, 1888, 676–678.
40. Gürich, G.: Die Kupfererzlagerstätte von Wernersdorf bei Radowenz in Böhmen. Zeitsch. für prakt. Geol., 1893, 370–371.

Miscellaneous.

41. Bell, J. Mackintosh: The Spassky and Atbasar Copper Mines in Siberia. Min. and Sci. Press, vol. CXX, 1920, 825–829.
42. Becher, H. M.: On Some Cupriferous Shales in the Province of Honpeh, China. Jour. Geol. Soc. London, vol. XLII, 1886, 494–495.
43. Voit, F. W.: Das Kupfererzvorkommen bei Senze do Itombe in der portugiesischen Provinz Angola, Westafrika. Zeitsch. für Prakt. Geol., 1902, 353–357.

PEGMATITES.

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FOREWORD.

THE pegmatites are undoubtedly the most bizarre, the most contradictory, the most complex and altogether the most interesting group of rocks known. To the mineralogist they furnish a long list of minerals, many of which are rare and are found in no other rocks, and many of the minerals are so beautifully colored that one need not be an adept to admire them. The professional geologist with many years of experience in specimen hunting is complacent indeed if the pegmatites do not even yet bait him with enticing problems.

They contain minerals that are deposited at temperatures near that of molten granite and they carry minerals that are deposited by comparatively cool water. They may be as simple in composition as a granite, or a list of the minerals from a complex pegmatite may look like a page from the index to Dana's System of Mineralogy. Geologists studying their origin have called them dikes, others have called them veins, and still others have found them to be both.

Naturally such an attractive and involved subject has given rise to a large literature, so extensive that one who reviews it with any thoroughness must have tremendous energy and enthusiasm and much time.

GENERAL FEATURES.

Origin of Name.—Kemp quotes A. Brogniart as giving Hauy credit for introducing the name pegmatite sometime before 1813, but at that time it was apparently applied only to graphic granite, a perfectly intelligible use of the word as derived from the Greek "pegma" (πῆγμα) meaning something fastened together, in a fanciful allusion to the pegging or fastening together of the

feldspar by rods of quartz. Graphic granite, however, was only one constituent of a disorderly assemblage of coarsely crystallized minerals, and the name was soon applied to the whole mass.

When less was known about them, pegmatites were much easier to define, so that the Century Dictionary's definition "A coarsely crystallized granite," or the older, more sententious "Giant granite" might have been acceptable. But many pegmatites do not resemble granites either in structure or composition, though nearly all accompany the more siliceous types of granular igneous rocks. Pegmatites accompanying the more basic rocks, especially rocks as basic as the amphibolites, pyroxenites or gabbros, are rare.

Occurrence.—To put it in another way, pegmatites are found only with those plutonic rocks that form wet melts and whose freezing is accompanied by the expulsion of watery solutions, gaseous or liquid—the same rocks which under proper conditions form tactites (contact-metamorphosed rocks). Pegmatites are found in the outer shell of the granitic rock but mostly in the adjacent rocks and are probably associated only with the domes in which are located the drainage systems of freezing batholiths. No pegmatites are known to originate from lavas; they are never found in unmetamorphosed sediments or lavas.

In shape and size they may range up to 250 feet or more thick and a mile or more long. They may form lenses or oval masses as much as 200 by 300 feet in cross section; "eyes" or pipes from a few inches to more than a hundred feet across.

In texture and mineralogy the differences are even greater. The minerals may be mostly small, some may be in huge crystals. The mineralogic composition may include little more than feldspar, quartz, muscovite and a little black tourmaline; the mass may be mostly quartz, or feldspar, or some one rare mineral like cryolite; or one pegmatite, or a part of it, may contain scores of minerals, many of which are rare, and the mineralogy may change with the rocks traversed. Two things they have in common—coarseness of texture and an extremely uneven segregation of minerals.

The distribution of the pegmatites is of course confined to areas underlain by plutonic rocks. The great number of pegmatites in some areas is astonishing, but quite as astonishing are the great lengths of granitoid intrusions in the Sierra Nevadas, the Andes, the Pacific coast ranges, the Rocky Mountains, and others, that show only insignificant pegmatites or none at all.

Economic Minerals of the Pegmatites.—A number of economic minerals are obtained only from pegmatites. These include massive feldspars of both potash and soda varieties, large clear quartz crystals, cryolite, columbium and tantalum minerals, beryllium minerals, gem tourmalines, sheet mica, primary lithium, rubidium and cesium minerals, uraninite and yttria minerals.

Minerals found both in pegmatites and in other types of deep-seated deposits include monazite, cassiterite, apatite, zircon and the andalusite-sillimanite-kyanite group of minerals.

Minerals of the middle and upper vein zones, but also found in pegmatites, are arsenic, bismuth and copper minerals, fluorspar, galena, gold, magnetite, molybdenite, sphalerite, scheelite and wolframite. This third group of minerals has provided one of the great stumbling-blocks in the understanding of pegmatites, for it has seemed necessary to most students either to make the temperature at which the pegmatite was formed low enough to allow the deposition of the comparatively low-temperature minerals, or else do what seemed more logical and admit that the pegmatite had been formed at a high temperature and then believe that low temperature minerals also had been deposited at the same high temperature.

TYPES OF PEGMATITES.

Like most other knowledge, our understanding of pegmatites has grown slowly. Their connection with igneous rocks has been readily apparent, as has been the family resemblance between the various pegmatites.

Owing to this family resemblance a like origin has usually been assigned to all, and until recently all were considered to be dikes or veins whose formation was accompanied by a great quantity of "mineralizers," including hydroxyl, fluorine, chlorine, phos-

phorus, boron, sulphur, and lithium, which were supposed to allow or promote the formation of large, even huge crystals, the substance for which was contained in the particular mass of magma which formed first the molten, and later the frozen, dike or vein. Only in the last few years has it been recognized that the origins of different types of pegmatites have been various.

According to their origin, pegmatites may be divided into three main classes and a number of subclasses:

1. Dikes
   - *a* Granitic
   - *b* Feldspathic
2. Veins
   - *a* In insoluble rocks
   - *b* In soluble rocks
3. Metamorphic pegmatites
   - *a* Progressively replaced dikes
     Subsequent quartz veins
   - *b* Pipes

Pegmatite Dikes.

Granitic Pegmatite Dikes.—Granitic pegmatite dikes are made up of distinctly granitoid minerals—quartz, potash feldspars, and in most, muscovite, in others biotite or both micas. Usually subsidiary black tourmaline, and in many dikes, garnet, are present, but both are probably always replacement minerals.

Although the mineralogy of the dikes is close to that of the granite, the minerals are of irregular size and distribution. The individual feldspars may reach a foot in thickness, but if they do they almost invariably inclose other minerals, and they are not commonly equidimensional, as in normal granites. The minerals do not reach spectacular sizes, nor do the dikes contain unusual minerals. Such pegmatites resemble greatly some of the coarse uneven granites evidently from the top of a wet granite intrusion, which look as if they had been stirred by convection currents in a boiling magma—probably due in part to movement while only partly solidified but in greater part to recrystallization under the

comparatively great quantity of fluids present and flowing through them. Many of the granites now used as building stone are of this type, and in fact such granites are frequently referred to as pegmatitic.

The granitic pegmatite dikes may range in width from a few inches to many feet and the length may reach more than a mile.

As already intimated, these dikes are intruded as very wet magma, and the freezing of the dike and the expulsion of the fluids are followed or accompanied by flow of fluids from the main magma so that more or less replacement or automorphism takes place with the introduction of pyrite, garnet, mica, black tourmaline, and other minerals, first along rather restricted channels and then more broadly. It is difficult to tell what part of many of these dikes may be original and what part may be automorphosed, but parts may be so changed as to show little resemblance to the original dike and they leave no doubt of their varied history.

Feldspar Pegmatite Dikes.—Certain other pegmatites, composed almost wholly of perthitic potash feldspar, also seem to be dikes. They appear to have been formed by the intrusion of a segregated part of the magma especially rich in feldspar, and very wet, so that large crystal growth takes place. Such dikes may be 200 feet wide and a half-mile long, with mixtures of the granitic minerals along the edges. Usually graphic granite, black tourmaline, quartz and a little muscovite show the work of later solutions.

Graphic granite, divided into variously oriented blocks a foot or more across, makes up similar dikes and seems to represent the work of after-solutions which carried practically nothing but quartz. At Winterham, Va., is a pegmatite up to 12 feet wide composed mostly of amazonite, the green potash feldspar, plainly bleached and replaced along cracks by other minerals. The Bennett pegmatite near Buckfield, Me., is perthitic feldspar about 40 feet wide with a working face 60 feet or more high. One end of the dike has been thoroughly replaced. At another place the start of graphic quartz intergrowths can be seen extending out-

ward from a quartz vein. In the Minor Keith pegmatite near Auburn, Me., the production of graphic granite from a perthitic feldspar dike by the introduction of quartz from quartz veins is almost diagrammatically shown.

Pegmatite Veins.

Pegmatite Veins in Insoluble Rocks.—Some narrow tabular pegmatites, particularly those lying along the bedding planes of comparatively insoluble rocks, such as are most schists, are made up of quartz and feldspar in more or less deformed crystals normal to the walls, with small plates of muscovite or biotite, some black tourmaline and other minerals. The middle may be vuggy and both feldspar and quartz may show crystal terminations. Such pegmatites are apparently veins.

But pegmatites of this general type vary greatly in the complexity of their mineral composition and may carry bismuth, chalcopyrite, molybdenite, pyrite, scheelite, and other minerals, as do ordinary quartz veins. A crust one-half inch thick from such a pegmatite vein near Washington, Ariz., contained potash feldspar, scheelite and quartz, all of which showed crystal faces.

In the Natas mine in middle South-West Africa such a vein in schists carries crystals of scheelite more than four inches thick and weighing more than 11 pounds; small crystals of orthoclase and some albite are along the walls. The center is filled with quartz.[1]

The vein also carries biotite, a little muscovite, chalcopyrite, bornite, chalcocite, molybdenite, a little visible gold in the quartz, a little apatite, tourmaline and specular hematite.

In addition to the minerals named, other veins of the vicinity carry calcite and some siderite.

The scheelite seems to indicate that the veins were deposited at rather low temperature, but the tourmaline points to a period of somewhat higher temperatures.

[1] Reuning, E.: Die Natasmine in Südwest-Afrika, eine pegmatitisch-pneumatolytisch-hydrothermale Übergangslagerstätte mit Scheelit, Molybdänglanz, Kupfererzen und Gold. Neues Jahrb. f. Mineralogie, vol. 52, pp. 192–264, 1925.

At Caceres, Spain,[2] a vein in mica schists carries amblygonite, cassiterite, apatite and quartz, and judging from the contained minerals, was apparently deposited at higher temperatures than the Natas vein and has decidedly pegmatitic characteristics.

Veins in more soluble rocks are very irregular and usually carry minerals formed by reaction of the solutions with the wall rocks. The unique uranium-bearing veins at Chinkolobwe, Katanga, in the Belgian Congo, have been called pegmatite dikes by Rickard,[3] who says that they cut across the copper deposits. They [4] swell and pinch like a rosary, cutting red schists, talcose rocks, and siliceous rocks which are dolomitic in depth. The veins average 8 to 12 inches, but reach 4 ft. in width. The vein matter was principally uraninite in aggregates of cubic crystals now mostly altered to torbernite near the wall and to gummite and uranophane in the center of the vein. The veins also carry quartz, feldspar, heterogenite,[5] molybdenite,[6] gold, selenium-bearing linneite,[7] slightly auriferous pyrite, arsenopyrite and specular hematite. Blocks of mineralized vein matter contain as much as 106 cu. ft. and weigh more than 22 short tons.[8]

The Canadian phlogopite deposits have long been a puzzle over which diverse views have been and still are held. They are found in rocks of the Grenville series of schists and limestones, through a belt of territory 50 miles broad extending from Kingston,

[2] Dorpinghaus, W. T.: Amblygonite-Zinn Vorkommen von Caceres in Spanien. Ein neuer Typus pneumatolytischer Lagerstätten. Archiv. für Lagerstätten-Forschung, vol. 16, pp. 1–49, 1914.

[3] Rickard, T. A.: A Journey to South Africa. Eng. & Min. Jour. Press, vol. 121, p. 19, 1926.

[4] Buttgenbach, H.: Note préliminaire sur des Minérais d'Uranium et de Radium trouvés au Katanga. Soc. géol. Belgique, vol. 44, 55 pp., 1921. Mineralogie du Congo Belge, p. 62, 1925.

[5] Schoep, Alfred: Les Mineraux Uraniferes (radioactifs) du Congo Belge, becquerelite, curite, soddite, chinkolobwite, kasolite, dewindtite et parsonsite. Soc. Belge Geol., Pal., Hydrol. Bull., vol. 33, pp. 169–192, 1924.

[6] Van Aubel, René: Sur quelques Minéraux et Roches du Haut Katanga. 3rd note. Soc. Geol. Belgique Ann. Pub. rel. Congo Belge Sup. p. C 71, 1928.

[7] Schoep, Alfred: Présence d'Or Natif et de Linnéite Sélénifere dans la Pechblende du Katanga. Soc. Geol. Belgique Ann. Pub. rel. Congo Belge Sup. pp. C 27–C 28, 1926.

[8] Buttgenbach, H.: *Op. cit.*

Ontario, 150 miles northeast into Quebec. The rocks contain many bodies of phlogopite forming veins, pockets and irregular masses which are in or alongside of masses of pyroxene. A few are found with aplite in white limestone. Large quantities of apatite are associated with the phlogopite and a number of other minerals common to the pegmatites also are found.[9]

The phlogopite deposits in Madagascar,[10] a country noted for its pegmatites, and those in Russia are similar to those of Canada. The deposits are apparently formed by pegmatite veins in the easily soluble limestones, and the close relationship to contact-metamorphic deposits is striking. In the Grenville limestone large bodies of microcline are also deposited from pegmatite veins.[11]

The Metamorphic Pegmatites.

The most complex of the pegmatites remain to be considered. They are of the type furnishing the cryolite of Greenland, the lithium, cesium, beryllium and other rare minerals. Many carry great numbers of minerals, others comparatively few. The minerals may form great masses or huge individuals. "Giant granite" describes part of them in that they carry the same minerals as granite. Beyond that the resemblance fails. Some of the huge crystals may be worth mentioning even though in part they are well known.

A book of muscovite from an undesignated mine in North Carolina is said by Kerr[12] to have made "two full two-horse wagon loads." An irregular book from the Pink mine was about 2½ by 5 feet and was 2½ feet thick, but it was not all good mica.[13] This year a book 3 by 3½ feet and 32 inches thick, weighing 4,320 pounds, was taken from the Spruce Pine Mica Company's No. 10 mine at Spruce Pine and was all merchantable

[9] Spence, Hugh S.: Mica. Canada Mines Branch, No. 701, pp. 37–75, 1929.

[10] Lacroix, A.: Mineralogie de Madagascar, vol. II, pp. 145–147, 1922.

[11] Shaub, B. M.: A Unique Feldspar Deposit near DeKalb Junction, New York, ECON. GEOL., vol. 24, pp. 68–69, 1929.

[12] Kerr, W. C.: The Mica Mines of North Carolina. A. I. M. E. Trans., vol. 8, p. 459, 1880.

[13] Horton, F. W.: Personal communication.

mica.[14] A crystal of phlogopite from Ontario, exhibited at the St. Louis Exposition, was more than 5 feet in diameter.[15] A book of biotite 6 feet wide, 9 feet long and 2 feet thick was found on the farm Hella, Stokken district, near Arendal, Norway.[16]

Aggregates of mica are, of course, much larger. A mass of amblygonite in the Peerless mine at Keystone, S. Dak., had been exposed in 1908 for a length of 40 feet, a width of 20 feet and 20 feet deep at one end. It yielded 900 tons of amblygonite. A regularly crystallized beryl on the Ingersoll claim 2½ miles northwest of Keystone, S. Dak., was 45 inches across the end between sides, but of unknown length, and a mass of columbite containing some quartz weighed a thousand pounds.

Many of the complex pegmatites, such as those at Baringer Hill, near Llano, Tex.; Mount Mica, at Paris, Maine; Pala, Calif.; and Brazilian localities, contain vugs in some of which a man could stand, which may contain quartz crystals weighing a thousand pounds apiece. In the United States National Museum is a flawless quartz ball 13¼ inches in diameter cut from such a Burmese crystal.

In the Etta pegmatite near Keystone, S. Dak., a single crystal of spodumene was $42 \times 6 \times 3$ ft. weighing about 65 tons; a mass of pollucite in the Tin Mountain pegmatite was 6 feet across. But the giant of giants was a perthitic microcline crystal mined on the farm Tveit, Iveland district, north of Kristiansand. It was $7 \times 12 \times 30$ feet and left a mold from which the crystal form could be determined.[17]

Wholly unique among the pegmatites is the cryolite mass at Ivigtut, Greenland, which is measured in hundreds of feet.[18]

[14] Urban, H. M.: Mica Mining Methods, Costs and Recoveries at No. 10 and No. 21 Mines of the Spruce Pine Mica Co., Spruce Pine, N. C. U. S. Bur. Mines, Inf. Circ. 6616, p. 4, 1932.

[15] Cirkel, Fritz: Mica, its Occurrence, Exploitation and Use. Mines Branch, Canada, p. 75, 1905.

[16] Andersen, Olaf, former Director, Norwegian Geological Survey: Personal letter, Feb. 18, 1932.

[17] *Idem.*

[18] Ball, Sydney H.: The Mineral Resources of Greenland. Meddelelser om Grønland, vol. 63, pp. 17–31, 1922.

If cross-sections of the complex pegmatites are taken at various levels they are usually very different, as in most replacement bodies. In the simpler metamorphic pegmatites, great masses of quartz are unevenly segregated. Large mica books, if present, are commonly within 4 or 5 feet of the walls, but in thick pegmatites with old cracks, mica books may be found ranged along the cracks in the central body, whence arises the report that the mica books "are found all through the pegmatite."

The Etta pegmatite has been mined to a depth of several hundred feet and there has been opportunity to see the changes of composition much better than in most pegmatites. It formed the point of a hill and originally showed at the top a rounded mass about 200 by 250 feet, inclosed in a gray quartz-mica schist. Occupying most of the middle half were large crystals of spodumene buried in feldspar and quartz and lying in every possible direction. Surrounding this central part irregularly were aggregates of various minerals—a mixture of quartz and microcline in which was muscovite in plates 4 to 6 inches across; a fine grained yellow mica forming solid masses with inconspicuous albite; spongy masses of honey-yellow mica in plates ⅛ inch to ¾ inch wide with small crystals of albite and cassiterite; blue apatite and other phosphates; irregular masses of white beryl; columbite enclosed by albite, quartz, beryl or honey-yellow mica; arsenopyrite in small masses and long needles; here and there glassy quartz in vein-like form carrying chalcocite and stannite. Spodumene in smaller crystals extended to the wall.

As the top was removed there were exposed great groups of spodumene crystals, and some that are more than 6 ft. long are radial in three dimensions from a dirty, incongruous, comparatively fine-grained mass 1 to 4 feet across. These central masses are worth noting for they are like nothing else in the pegmatite. They have never been studied in detail, but they contain apatite, triplite, muscovite in plates an inch across, some quartz, potash feldspar and other minerals, and are replaced masses of country rock.

Lower down, quartz has become more prominent at the expense

of feldspar and the micas, the cassiterite and columbite have nearly disappeared and beryl is also apparently in much less quantity.

In the Hugo pegmatite, a short distance northwest from the Etta, and possibly 150 ft. across, great masses of amblygonite and large spodumene crystals occurring in the upper part are succeeded below by microcline, and a couple of hundred feet below the outcrop a drift has cut a granitic pegmatite dike three or four feet thick, which is apparently the lower part of the Hugo.

The Peerless pegmatite forms the crest of a hill south of Keystone. In plan it is somewhat like the horizontal cross section of a tadpole—a fat body and a slim tail. It is about 250 feet long and possibly 150 feet across. The top is said to have stood 30 feet above the hill and to have been mostly quartz. I have already spoken of the great mass of amblygonite that yielded 900 tons. No other lithium minerals have been seen in the Peerless, and after removing the large mass of amblygonite little more of the mineral was found. The great quartz mass apparently rested on a mixture of microcline, quartz, and muscovite. On the Ingersoll claim, in the upper part of the principal pegmatite, are big individual masses of potash feldspar, amblygonite, quartz and lepidolite. Fifty feet below there is only an impure mass of lepidolite.

At Camp Harding, New Mex., lepidolite inclosing some albite and quartz occupies from 10 to 25 feet of the underpart of a long, gently dipping pegmatite. Above is a sheet of quartz 5 to 15 feet thick containing spodumene crystals shaped like pieces of 6 inch boards, all pointing within 45° of vertical, and above that are sheets of quartz as much as 5 feet thick. The quartz-spodumene mass has been replaced by the lepidolite. Examples of such irregularity might be multiplied, and in fact irregularity is the regular thing.

The conception of replacement on a great scale in pegmatites has developed rather slowly. Replacement has been accepted readily in many contact metamorphic deposits because it was fairly obvious, but like the gap between a little gully cut by a

tiny trickling stream and a canyon like that of the Colorado River, the difference is so great that the mind is at first apt to see contrast rather than likeness of origin.

In the simpler metamorphic pegmatites, the replacement of part of the potash feldspar by quartz in the more or less regular forms called graphic granite is one of the first. Black tourmaline has invariably replaced earlier minerals, and is one of the first of the usurpers. Sheet mica is also one of the earlier replacement minerals. As I have said, it is near the walls, that is, in the cooler zone, or along cracks that had formed in an already solid body. In some pegmatites the mica known variously as feather, herringbone or "A" mica can be seen in all steps as the continued flow of solution has aided molecular readjustment and the mica has lost its blemishes and become, with or without increase in size, sheets of perfect muscovite. Muscovite in small books may take part in the replacement of pegmatite at any stage and place.

At one point on the Peerless claim, a vertical crack extended through the quartz core of the pegmatite and a mass of spongy yellow muscovite in plates 1 inch across by 1/8 inch thick with less albite, cassiterite and columbite, had replaced the quartz along the crack for 5 or 6 inches on each side, coming to a rounded end above, the crack continuing on in the quartz. Along the edge of the mica, a bristling border of albite crystals jutted into the quartz. Some columbite crystals extended beyond the albite, and irregular small aggregates of columbite and cassiterite occurred between the mica books. A well-formed crystal of columbite at the edge of the mica, but largely imbedded in the quartz, weighed more than 200 pounds. It is patent that the quartz had been replaced by minerals brought in by solutions following the crack. Columbite penetrated farthest, followed by albite and muscovite. At another place in the same quartz mass the terminal face of a beryl crystal 19 inches across was bordered for half of its periphery by a halo of albite 9 to 13 inches wide, and outside of that was a row of honey-yellow muscovite plates 3/4 inch thick. On one side of the beryl was an aggregate of ordinary nearly white mus-

covite. Thus in the same mass of quartz at one place albite preceded mica in the replacement of the quartz, and 75 feet away the succession was reversed. Where the quartz was brecciated albite had followed along the cracks and made noticeable replacements, as is common in other pegmatites.

In the great potash feldspar pegmatite on Crabtree Creek, 7 or 8 miles south of Spruce Pine, walls perhaps 100 by 300 feet have been left by quarrying and some of the replacement processes are diagrammatically shown. The edges of quartz veins run diagonally through the feldspar, and on either side of the veins large crystals of feldspar have been replaced by radial ribbons of muscovite, extending only to the edges of the individual crystal.

A specimen of beryl from the Old Mike mine near Custer, South Dakota, shows beryl replacing potash feldspar, and similar specimens come from Latah, Idaho. In each, the beryl forms a shell the interior of which is still feldspar. In a specimen from the Hugo mine, beryl is being replaced by albite. A beryl from Granite, Colo., is largely replaced by muscovite.

In places quartz is replaced by lepidolite and albite, and vice versa, lepidolite is replaced by quartz. In a specimen from Camp Harding, New Mexico, lepidolite and albite may be seen in waves replacing microcline. Amblygonite and spodumene replace microcline and quartz. Pollucite and the uranium and rare earth minerals apparently replace most earlier minerals.

In general, black tourmaline forms on and near the wall of the pegmatite, garnet near the wall (and any crack in the frozen pegmatite may act as a wall), large books of muscovite and beryl next, and the other minerals named form in that part of the pegmatite which becomes hottest under the flow of invading solutions. Amblygonite and spodumene form earlier than lepidolite, cassiterite, the rare earth, uranium and columbium-tantalum minerals, and most albite.

As soon as most minerals have been fairly well formed, new ones have been brought in, or the old ones have been wantonly destroyed. Thus tons of pollucite in Tin Mountain were reduced

to clay, and in Maine, clay or cavities were left. Pollucite is also replaced by lepidolite. Amblygonite is replaced by albite, lepidolite, apatite, uraninite and other minerals, but few minerals replace spodumene. However, like pollucite, it is altered to a clay-like mass. Quartz is omnipresent and apparently may replace most minerals. In some pegmatites fluorspar, galena, wolframite, molybdenite and other minerals of the low temperature veins are found replacing earlier minerals.

I can see only one thing that can explain all these changes—a continued flow of solutions that progressively replace the earlier constituents of the dike. If a coarse wet granite is intruded into the rocks it is chilled. If the magma is very wet, fluids continue to flow through the mass for a long time and are apt to follow the cracks into which the dikes were intruded. The fluids first heat the dike and successively replace the minerals by those of higher temperature.

If they flow long enough, a time must come when the solutions cool and carry minerals deposited at lower temperature along the same paths, where mass action and probably the continued presence of fluorine cause them to replace the minerals deposited at higher temperatures, or fill vugs formed by other solutions.

Metamorphic pegmatites probably show by their less common constituents derivation from different magmas. The pegmatites of the Piedmont section of the Carolinas carry cassiterite and purpurite (an alteration mineral from lithiophylite); in the mountains both north and south of Spruce Pine, N. C., uraninite and other uranium minerals, rare earth and tantalum-columbium minerals are found as accessories in the pegmatites; and in the area near Tuxedo (Zirconia P. O.) a third group of pegmatites carries zircon, sphene and iolite. These lesser minerals seem to indicate three different magmas as sources of the respective groups of pegmatites.

The solutions are probably very dilute and the length of time that they flowed to deposit feldspar and cryolite by the hundred-thousand tons, lithium minerals by the thousand tons and pollucite by the tens of tons, must have been great. However, there has

been plenty of time. Uraninite from the Ingersoll No. 3 dike showed an age of almost 1,700,000,000 years.[19]

Flow of solutions may stop and begin again after long periods. A samarskite from Petaca, N. Mex., showed a difference in structure and radioactivity, and analysis showed that a core had been replaced 150 million years after the older part was formed.[20]

Other strange freaks take place. At Tin Mountain a biotite that carries 3½ per cent cesia has been deposited at the edge of the pegmatite,[21] although the pollucite and cesia-bearing lepidolite occupy the hottest places. On Hodgeon Hill, near Buckfield, Maine, a complex metamorphic pegmatite has replaced a granite dike, here partly, there wholly, through its width of 15 feet. In some granitic pegmatite dikes, a pocket only has been replaced by quartz, spodumene and other minerals. The Tin Mountain pegmatite with its great masses of rare minerals is the replaced end of a granitic pegmatite dike more than half a mile long. At the other end is a smaller metamorphosed part of the dike. Other dikes of the area are also metamorphosed in parts only.

In some places, as in the Deep Creek Mts. of Utah, in South Africa,[22] and in New South Wales,[23] granites are metamorphosed in pipes or eyes in much the same way and are also to be classed as pegmatites. They differ in form only from the pegmatite just mentioned as occurring at Hodgeon Hill, near Buckfield, Me. Many of the pegmatites between Spruce Pine and Penland, N. C., have apparently been formed in the same way, that is by solutions following cracks and gradually replacing the granite with large pegmatitic minerals.

[19] Davis, C. W.: Composition and Age of Uranium Minerals from Katanga, South Dakota, and Utah. Amer. Jour. Sci. (5), vol. 11, p. 215, 1926.

[20] Hess, Frank L. and Wells, Roger C.: Samarskite from Petaca, New Mexico. Amer. Jour. Sci. (5), vol. 19, pp. 18–26, 1930.

[21] Hess, Frank L. and Fahey, J. J.: Cesium Biotite from Custer County, South Dakota. Amer. Miner., vol. 17, No. 5, pp. 173–176, May, 1932.

[22] Wagner, Percy A.: The Mutue Fides-Stavoren Tinfields. Union of South Africa Geol. Surv. Mem. 16, 1921.

[23] Andrews, E. C.: The Geology of the New England Plateau, with Special Reference to the Granites of Northern New England. New South Wales Geol. Survey Rec., vol. 8, pp. 108–152, 1909.

And now perhaps I may offer a definition of pegmatite:

Pegmatite: A general name for rocks with coarsely and unevenly crystallized and segregated minerals occurring as dikes, veins or metamorphic masses formed from the aqueous solutions of a freezing magma or from the combination of the solutions with previously existing minerals.

U. S. Bureau of Mines,
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THE MIAMI–PICHER ZINC–LEAD DISTRICT.

W. A. TARR.

Two papers relating to zinc-lead ores have recently appeared, one by Weidman [1] on the ore deposits of the Miami-Picher zinc-lead district, and the other by G. M. Fowler and J. P. Lyden,[2] on the ore-deposits of the Tri-State (Missouri-Kansas-Oklahoma) district. These two papers present the results of studies extending over a period of several years and are especially important contributions to the literature on the deposits because they give details of the more recently developed Oklahoma portion of the district.

The authors of both papers advocate a magmatic source for the zinc-lead mineralization, which adds their names to the steadily growing list of geologists (myself included) who favor such an origin. I regret that so little space was devoted by these writers to the composition of the mineralizing solutions, the localization of the deposit, and the method and cause of precipitation, but this was probably due to the fact that the two papers are, in some measure at least, preliminary or progress reports. It is to be hoped that subsequent papers will cover fully these important points.

This article is written to take issue with these authors on their interpretation of the age and origin of the chert; also, to point out that to regard this chert as an original chemical deposit, formed simultaneously with the limestone of the Boone formation, makes possible a clearer, more logical, and more accurate account of the origin of the ores. The discussion will be in the following order: first, the origin of the chert (not the jasperoid), second,

[1] Weidman, Samuel. The Miami-Picher Zinc-Lead District, Oklahoma. Okla. Geol. Surv. Bull. 56, 1932. University Press, Norman, Oklahoma.

[2] Fowler, G. M., and Lyden, J. P. The Ore Deposits of the Tri-State District (Missouri-Kansas-Oklahoma) (with Discussion). A. I. M. M. E., Tech. Pub. 446–I. 33., Jan., 1932.

the origin of the chert breccias associated with the ores, and, third, the origin of the jasperoid and zinc-lead ores.

ORIGIN OF THE CHERT.

Many critical studies of chert have been made in recent years, and although the investigators are not in full accord, there is an unmistakable trend toward regarding chert as being a direct chemical precipitate of colloidal silica formed at the time of the deposition of the calcareous sediments with which chert is so commonly associated. The writer published, in 1917,[3] his first paper on this subject, in which for the chert in the Burlington limestone (called the Boone formation in the Tri-State district) of central Missouri he advocated an origin by direct chemical precipitation on the sea floor.

Since 1917, the writer has extended his study of chert to that of other states and abroad. The results of these later studies have amply supported his earlier view. Other investigators have applied the theory to other deposits of chert and have usually found it applicable. During the past year, the writer, in conjunction with W. H. Twenhofel, prepared for the revised second edition (1932) of "A Treatise on Sedimentation," the part relating to the origin of chert and flint, in which are given the details of the source of the silica, the method of its precipitation, the origin of the nodules and beds, and the subsequent changes from colloidal silica to chert and flint. It is sufficient here to state that the quantity of silica being carried to the sea annually ranks next to that of calcium. In the past, physiographic conditions have favored an even greater supply of silica. At such times, the dominant sediments deposited were calcareous and siliceous, the latter a chemical precipitate. Organic silica was long regarded as the source of the silica in chert and flint, but microscopic studies and studies of insoluble residues have shown that such a source is wholly inadequate.

[3] Tarr, W. A.: Origin of the Chert in the Burlington Limestone. Am. Jour. Sci., vol. 46, 1917, pp. 409–452. See also a later study, Tarr, W. A.: Origin of Chert and Flint. Univ. of Missouri Studies, vol. 1, no. 1, 1926.

Weidman estimates that the primary silica (both clastic and organic) of the Boone formation averages about 10%, and he recognizes that this amount is wholly inadequate to account for the enormous amount of chert in the formation. From a study of the insoluble residue (mainly chert) of the Ottawa County well to the south of the mining district, he concludes that there is on the average 56% of chert present in the Boone formation, and in the adjacent Tri-State mining district he finds about 66%. To account for these large amounts of chert in excess of the original 10% of silica, Weidman postulates a pre-mineralization period of silicification to furnish the extra 46% of the Boone formation in general, and believes that the final 10% in the mining district came in with the ores.

Fowler and Lyden recognize that "minute quantities of silica were deposited with this limestone (Boone)," but state that outside of the disturbed areas "little or no chert is found." They regard the Boone formation outside the Tri-State district as being free from chert, a conclusion that shows they are unfamiliar with the formation elsewhere. Weidman does not clearly give his idea of the composition of the Boone formation outside the area he was studying. Some of his statements seem to indicate that he recognized it as being cherty elsewhere, since he states: "The Boone formation named from Boone County, Arkansas, its type locality, is a cherty limestone formation,"[4] and further, —"The Boone, as determined from samples of the deep well of the Ottawa County Welfare Home outside the mining district . . . , is a limestone formation containing a variable but usually large content of flint or chert."[5] The insoluble residue (mainly quartz as chert or flint) of the Boone formation ranges from 16 to 80%, hence Weidman rightly states that the usual name "Boone chert" is a better name for the formation than "Boone limestone."

It should be stated that the Boone formation (or "Burlington" as it is more commonly known) is extremely cherty throughout its distribution in Missouri, Oklahoma, and Arkansas. One

[4] *Op. cit.*, p. 11.
[5] *Idem*, p. 13.

needs only to note casually its outcrops to see this. The chert in this formation, as the writer has shown in the articles cited, is primary, having been deposited with the limestone. Its distribution over thousands of square miles, the widespread occurrence of the chert in the same zone, its persistence along a given horizon within that zone, and innumerable other lines of evidence, prove its syngenetic origin. Fowler and Lyden state that "the chert nodules have the same characteristics in each bed over many square miles," a fact which the writer has proved for many chert and flint deposits in widely scattered areas. Fowler and Lyden raise the question of the origin of these chert nodules and the cause of their shape and size, and ascribe these features to some inherent property of the limestone. The illustrations they give of chert nodules in the limestone (which has not been metamorphosed, as they state) could be used to illustrate the occurrence and form of chert in hundreds of other localities. The limestone had no influence upon the form or other features of the chert nodules. Colloidal silica tends always to form rounded masses, as surface tension is the chief controlling force, and so when precipitated on the sea floor it assumes this form. The dominantly elliptical form of chert nodules is the result of flattening caused either by the weight of the colloidal mass itself when it becomes large (just as mercury and large drops of water are flattened) or by the weight of the calcareous muds deposited above it. Where the quantity of colloidal silica is large or deposition is rapid, a bed or layer of chert results. These layers commonly range in thickness from less than one inch to several feet, but beds that are 10, 20, 50, or more feet thick are known. Attempting to conceive of the secondary origin of such beds is like explaining a bed of limestone as secondary in origin. It cannot be done.

In discussing the paper by Fowler and Lyden, C. L. Dake[6] argues that all chert is a superficial phenomenon and that it does not occur at any distance from the outcrop. This statement has been made before and is so obviously refuted by the occurrence

[6] Fowler and Lyden, *op. cit.*, pp. 39–41.

of chert in thousands of wells at all distances from the outcrops that it scarcely deserves mention here. Dake makes the surprising statement that it would be impossible to drill through solid masses of chert, apparently unaware that for years miners have drilled quartz, and well-drillers, chert.

In connection with the occurrence of chert and flint, the studies of insoluble residues made by the Missouri Bureau of Geology and Mines are often cited as proving that chert is not present, but only disseminated silica that later forms chert. *Chert is present*, however, and in quantities comparable to the amount exhibited by the formation at its outcrop. The writer is only one of hundreds (including McQueen and others who use insoluble residues) who have studied well cuttings and recognized the various cherty formations by the large quantity of chert in the cuttings. McQueen's log of the Bird Dog well, given by Fowler and Lyden (p. 17), is sufficient proof of the abundance of chert. Weidman's figures are even more startling for the same well, for he gives the chert residue as ranging from 19 to 97% and averaging about 62%. This sort of data regarding chert in well cuttings proves that chert is found in the same quantity and manner far from the outcrop as well as near it, and also that it occurs unassociated with unconformities. All this evidence is favorable to the view that chert and flint are original primary constituents of the sedimentary rocks and not secondary surficial products.

BEARING OF THE ORIGIN OF THE CHERT ON THE ORIGIN OF THE ORE DEPOSITS.

The significance of a syngenetic origin of the chert as applied to the lead-zinc deposits of the Tri-State district is two-fold. (1) The view of the authors of both papers that the chert was largely, if not wholly, introduced later, calls for the introduction of an enormous amount of silica. This introduction, they assumed (on wholly inadequate evidence) to be a pre-sulphide mineralization process, yet, as will be shown later, this is an absolutely unnecessary assumption. (2) If the chert is syngenetic, the events preceding the formation of the ore bodies fall

naturally into the geologic history of the area and are not necessarily a part of the mineralization process.

The View of Later Introduction.

In neither of the two papers do the authors raise a question as to the adequacy of the source of the silica. Weidman accepts the theory of the replacement of the limestone by silica, in spite of the character of the evidence advanced against this view in the papers cited above. He states (p. 65) that the silicification (chert formation) occurred during "a relatively short period in Middle Mississippian time," and holds that over 50% of the formation is chert and that fully 45% was introduced during silicification. He regards (p. 84) this silica as being of hydrothermal origin and therefore definitely places himself on record as believing that hot waters permeated these rocks in Middle Mississippian times. That the quantity of silica represented by the sedimentary chert of the Boone formation could be introduced from a magma without any accompanying sulphide formation or other type of mineralization represents a view that is opposed by a vast quantity of data regarding the deposition of quartz in mineralized areas everywhere. Weidman's further postulate of a succeeding period of silicification produced by magmatic solutions and accompanied by the formation of an abundance of sulphides seriously weakens his first postulate. There are strong reasons for not accepting his view about this early silicification. In no mineralized district where silicification of limestone from hydrothermal solutions has occurred do we find the silica assuming the form of isolated nodules and lenses. The process of replacement involves the entire formation. Innumerable facts, such as the distribution, both vertical and horizontal, of the chert in a formation and within a given bed, and the position of chert nodules along certain horizons, have not been adequately explained by authors favoring the replacement theory. These points are likewise involved in the hypothesis advanced by Weidman, and Fowler and Lyden, and are unexplained by them. The chemical precipitation of the chert (as indicated in the first part of this discussion) does

explain these facts of occurrence without requiring replacement. The hydrothermal theory does not account for the presence of the chert throughout the Mississippian formation at points that are far removed from the mineralized area. Neither could this theory be applied to cherty horizons of other periods and other localities. In short, why should one attempt to explain by a hypothetical mineralization process what is so evidently a normal process of sedimentation?

Fowler and Lyden regard all the chert (not jasperoid) as secondary in origin and ascribe it to a purely hypothetical (but wholly unnecessary) process called "metamorphism." The use of the term "metamorphism" in this sense is to be regretted. Their figure (Fig. 10) of a chert nodule in "a partly metamorphosed" bed would aptly illustrate thousands of widespread occurrences of chert or flint nodules in pure, unaltered limestones and even in chalk. They assume that the silicification (the term "chertification" is just another piece of excess baggage in our literature) accompanied the dolomitization or followed it, and state that the material for the two processes and for the formation of the sulphides probably came from the same original source. Since they regard the ores as being of magmatic origin, their assumption would necessarily be that the silica was magmatic in origin, also. The silicification, they assume, occurred only in association with disturbed areas, yet they comment on the similarity of nodules and beds of chert over wide areas throughout the district. It is evident that they recognize the syngenetic chert with its uniform features, and are confusing it with the irregular, brecciated masses of the same materials associated with the ores. Liesegang's theories regarding agate structures have no application to chert and flint deposits. As previously pointed out, Fowler and Lyden are in error in thinking that the Boone (Burlington) formation is not cherty outside the Tri-State district, and all the silica in the chert in this outside area certainly did not come from a magma.

The writer has seen the chert in all parts of this district and adjacent areas, and has not the slightest doubt that it was de-

posited along with the limestone. All the features of the chert are adequately and accurately explained under the theory of direct chemical precipitation. A discussion of the origin of the chert is, moreover, not necessary to an explanation of the origin of the ores if the chert is regarded as a chemical sediment. Of course, it is recognized that it is the prerogative of each to hold any view as to the origin of anything, but holding too consistently to views of ancient lineage is hardly progress.

Syngenetic Origin.

If one accepts the syngenetic origin of the chert, what is the sequence of events in the zinc-lead mineralization of the district?

Origin of the brecciated zones.—The original distribution of the chert was uniform along or within a given bed in the limestone. In some beds, the chert occurs as nodules, in others as elongated nodules or lenses, and it may also occur as beds that vary in thickness but are persistent over wide areas. Weathering during the interval between the Mississippian and Pennsylvanian periods liberated the chert, just as is being done today. This residual (not secondary) chert occurred in the soils, along the slopes, and in the stream gravels. The Pennsylvanian sea washed over and sorted these chert fragments and incorporated them in its lowest member as conglomeratic material. Locally, this conglomerate may have become the seat of suubsequent mineralization, but such occurrences are uncommon.[7]

Underground Circulation.—During this pre-Pennsylvanian erosion interval, an underground water system became well established, just as one has been established in the central Ozarks today, as is evidenced by the numerous large springs therein.

[7] C. K. Leith regards "the Great Falls chert as an old erosion surface, . . ." (ECON. GEOL., vol. 27, pp. 416–417, 1932), but this is an impossible concept, as the study of any exposed residual deposit shows. The Great Falls chert has none of the characteristics of a residual deposit and there is no residual clay with the chert breccia. It was not reworked on a beach (which would have removed the clay) for the fragments are sharply angular and unsorted. The fact that the *breccias follow runs and are not continuous throughout the Tri-State and adjacent areas* (as they must be if they represent an old erosion surface) is sufficient proof in itself of the untenableness of this concept.

In numerous quarries in the Burlington limestone, evidence of an underground system in the pre-Pennsylvanian period is indicated by old solution channels, now more or less filled with Pennsylvanian sediments, of which chert is an abundant constituent. We know very little about the land surface during this period, and so cannot make any deductions as to the direction of the underground water circulation. The movement of the underground water was facilitated by the presence of chert lenses and nodules but was especially active in the Grand Falls (Boone) chert horizon (about 30 to 35 feet thick). The solution channels followed this horizon chiefly, but also extended upward and downward. Their courses were typically curved yet were more or less irregular, just as one finds the courses of caverns today. If these solution channels were developed above the water table (as most modern caverns apparently have been), the physiographic conditions following the Mississippian period were different from those of today. The development of the channels (and accompanying breccias) required a long time, since solution and removal of material were slow, but, as the process could and undoubtedly did continue after burial under the Pennsylvania and possibly Mesozoic beds, there was ample time for their development. Emmons [8] regards the ores of this district as having been formed "about the time of the Appalachian revolution," but the writer sees no reason for assuming this date.

There is the further possibility that the development of this channel system could and possibly did take place below the water table, in the manner so ably presented by Prof. W. M. Davis.[9] Davis calls the channels that are essentially parallel to the bedding planes "galleries," and those openings that are more or less vertical, "shafts." The distribution of the ore bodies, which, of course, depict the original openings or brecciated areas in the Tri-State district (well shown on Plate II of Weidman's report) conforms remarkably to that of the galleries and shafts in hun-

[8] Emmons, W. H.: The Origin of the Sulphide Ores of the Mississippi Valley. ECON. GEOL., vol. 24, p. 221, 1929.

[9] Davis, W. M.: The Origin of Limestone Caverns. Bull. Geol. Soc. Amer., vol. 41, pp. 475–628, 1930; Science, vol. 73, pp. 327–331, 1931.

dreds of caverns. In fact, the study of Weidman's map, or of those in the Joplin folio, reveals such a striking similarity in the form of the caverns and of the ore bodies that one can scarcely doubt their very evident relationship. Another significant point in Davis' paper is that these caverns, made below the water table, may be lined with calcite crystals rather than dripstones (stalactites and stalagmites). Davis cites the crystal-lined caves of the district as examples of caverns formed and partially refilled below the water table. That such channels exist today was evidenced in drilling deep water wells on the campus of the University of Missouri, as the drill would drop 2 or 3 feet in cavities which were between 900 and 1,200 feet below the surface. Swinnerton [10] has questioned the adequacy of a deep-seated ground water to develop caverns, suggesting a "water-table drainage system" as more probable. Such a hypothesis may be applicable in some localities, but Swinnerton's arguments against Davis' theory of a two-cycle development are not sufficiently convincing to the writer to rule it out as a possible means of forming the brecciated zones in the Tri-State area. The writer believes that the various types of openings, now more or less filled with brecciated masses of chert, resulted when the limestone between and about the chert masses was removed by groundwater acting both above and below the water table.

Weidman (p. 42) rejects solution work as a means of forming the caves, and suggests that they were "caused by the arching of competent strata followed by subsequent settling underneath when the stresses are relieved," a far more improbable method, as it involves forces the source of which has been unaccounted for.

There is another possible means of dissolving a vast amount of calcium carbonate from the Boone formation and leaving a mass of broken chert behind. Those advocating a magmatic origin for the ores believe that igneous rocks have been in existence under this area for a considerable time, and, if this is true, warm water rising from the magma and passing through the Boone formation could have removed a vast quantity of calcium

[10] Swinnerton, A. C.: Origin of Limestone Caverns. Bull. Geol. Soc. Amer., vol. 43, pp. 663–693, 1932.

carbonate and redeposited it at the surface, as is being done today at Mammoth Hot Springs in Yellowstone Park. Subsequent erosion might very well have removed all evidences of such a deposit. This suggestion is in keeping with the possible relationship of the lead-zinc ores to a magma below.[11]

Breccia Formation.—Either of these methods (*i.e.* solution by ground water or hydrothermal solutions) would mean the removal of $CaCO_3$ and the concentration of chert. The presence of the chert, which has a weak contact with the limestone, would favor movement of water along a given cherty zone. If the limestone were removed uniformly, contraction would be uniform, but, for the most part, this was not so. The writer's studies of chert, especially of the larger nodules and lenses, have revealed that these silica aggregates are in a constant state of stress as they exist in the rocks. On the removal of the surrounding rock, this internal force becomes effective and aids in the mechanical disintegration of the chert mass.

Solutions in removing the surrounding limestone would loosen the chert mass which would then undergo disintegration, finally resulting in a *breccia* aggregate. It is through this process that the brecciated chert aggregates of the Tri-State and other districts were most probably formed. (The writer's notes show that he favored this method of origin of the breccia over twenty years ago, even before his intensive study of chert, and a similar method of origin has been advocated by others.) The settling of overlying portions of the formation would aid in further crushing and brecciating the chert. In the accounts of the occurrence of the ores, all writers note that the broken areas are irregular, horizontally and vertically; a distribution in keeping with the origin suggested here. The runs and circles (galleries) represent the courses of the underground water that were essentially horizontal. Where the water moved vertically, broken zones (shafts) developed along such courses. The removal of much limestone in a given area resulted in settling and disturbance of the continuity

[11] Since writing the above, the writer has seen a somewhat similar suggestion made by R. T. Walker (A. I. M. M. E., Tech. Pub. 154, 1928) in application to the deposits at Tintic, Utah.

of the individual horizons. The curved fractures that follow the runs (ore bodies) owe their courses to settling along, and adjacent to, these solution areas (plates 3 and 4, Weidman's report). Faulting had little to do with the localization of these galleries and shafts; jointing was probably a factor because it furnished divisional openings along which the water entered. Strong major joint systems were apparently absent, as little alignment or definite arrangement of the ore bodies is noted in the maps representing them. The writer does not regard the various tension shear zones, postulated by Fowler and Lyden, as at all essential. The inability of these men to suggest a satisfactory source for the force needed to produce the shear zones indicates the weakness of this concept. The origin of the breccias by solution is a much simpler explanation. Bain [12] undoubtedly thought solution work was a possible if not an essential factor even below the water table, but considered faulting as the more important cause. Weidman regards the breccias as being due to various causes, largely mechanical though supplemented by "chemical processes of mineralization including solution, deposition, and replacement." Weidman regards the Miami syncline (the "Miami shear trough" of Fowler and Lyden) as being due to compressive stresses caused by igneous intrusion, although in other statements he indicates that some faulting occurred. Fowler and Lyden evidently regard shearing (faulting?) as a major cause of this trough. According to them, solution work has accompanied deformation and probably induced local slumping. All these men regard the breccias as being dominantly of mechanical origin.

The distribution, both horizontal and vertical, and the character of the breccia materials (which include not only chert but undissolved limestone fragments) favor an origin due to solution rather than to mechanical movements. The disagreement of writers as to the location and origin of such mechanical movements casts strong doubt upon their existence at all. There is no reason, however, where faults occur in the district, why some

[12] Bain, H. F., Van Hise, C. R., and Adams, G. I.: Preliminary Report on the Lead and Zinc Deposits of the Ozark Region. U. S. Geol. Survey, 22d Ann. Rept., part II pp. 23–228, 1901.

brecciation should not be associated with them. The open spaces in these solution-formed brecciated masses furnished the areas for the deposition of the zinc-lead ores.

Before leaving the subject of the brecciated zones, it might be well to state that the writer regards the Seneca fault and the shorter faults having a general northeast-southwest strike as the principal faults of the area. The Miami syncline is probably a fault system (Weidman and Fowler and Lyden note faults along it) that roughly parallels the Seneca fault. The writer has found faults of unknown displacement paralleling these two in the area between them. These faulted zones are probably the channels through which the solutions rose, as noted below.

DEPOSITION OF THE ZINC-LEAD ORES.

The solutions (of magmatic origin, because the chemistry of any deposit containing much lead demands a transporting solution more efficient than any meteoric water not modified by magmatic waters) from below followed the faults upward until they reached the solution channels with their chert breccias, and, being unable to rise higher because of the Pennsylvanian shales above, they spread laterally along these channels. Both horizontal channels ("runs and circles") and the vertical brecciated zones were filled by these solutions, which were siliceous and alkaline. The silica was carried as a colloid and deposited as a fine-grained quartz (the "jasperoid" of the district) rather than as chalcedony. This fine-grained quartz is coarser grained than the chert. This fact is recognized by Weidman, and it seems strange that he failed to see that the two materials, the primary chert of sedimentary origin and the fine-grained quartz or jasperoid, were entirely different, and that the chert was not of hydrothermal origin. Had the colloidal silica of these last solutions been deposited rapidly, a colloform banded chalcedony (such as that described by Lindgren in the Tintic, Utah, deposits) would probably have resulted. Most of the jasperoid was deposited in the open spaces of the breccia, where it acted as a cementing material. Some of the silica replaced the limestone, although the abundance

of unfilled spaces in the ore bodies shows that not much replacement occurred. Silica deposition was accompanied by the deposition of much sphalerite (chiefly fine-grained) and some galena and dolomite.

For the most part, the dolomite in the ore solutions was deposited after the major precipitation of the silica, but the two minerals were also deposited simultaneously during the closing stages of the silica deposition. The deposition of sphalerite carried through the dolomite deposition, in places even exceeding it in amount, but the layers of large saddle-shaped, pink dolomite crystals mark the closing stage of dolomite deposition. More or less galena was deposited in this period. The deposition of the coarse, massive sphalerite overlapped and followed that of the dolomite, and was, in turn, overlapped and followed by deposition of galena. The closing stages of mineralization show mainly pyrite, marcasite, chalcopyrite, and a recurrence of sphalerite in small red crystals. Calcite is commonly later than these sulphides but may be deposited simultaneously with them, as the sulphide inclusions of much Joplin calcite show. All these minerals were deposited upon any of the preceding minerals.

This sequence shows that the main mineralization consisted of the first group of minerals: jasperoid, sphalerite, galena, and some dolomite, and that the sulphide deposition extended beyond that of the silica and dolomite. The solutions were capable of transporting silica, magnesium, zinc, and lead, as well as minor amounts of iron, copper, cadmium, and barium. Hydrogen sulphide, carbon dioxide, and probably HCl were present in the solutions. It should be noted that in the Tri-State district the rare metals germanium and gallium occur in the sphalerite,[13] and rare earths of the yttrium and cerium groups in the yellow calcite.[14] The sphalerite from Webb City, Missouri, was one of the five richest specimens examined by Urbain. The presence of cadmium and rare metals and earths in these deposits points to a magmatic source for the materials of the deposit.

[13] Urbain, G.: Comptes Rendus, vol. 149, p. 602, 1909.

[14] Headden, W. P.: Phosphorescent Calcite from Fort Collins, Colo. and Joplin, Mo. Amer. Jour. Sci., vol. 21, pp. 301–308, 1906. Pisani, F.: Comptes Rendus, vol. 158, p. 1121, 1914 (confirmed Headden's work).

The silica, as has been said, was transported dominantly as colloidal silica. Originally, some sodium silicate may have been in the solution but most of it would have hydrolized to colloidal silica and sodium hydroxide. The lead was undoubtedly carried as $PbCl_2$, as this salt is readily soluble in water at a temperature of 100° C. or above. The zinc was also probably transported as the chloride, though it may have been transported as some other compound, as other zinc salts are in general more soluble than the lead salts. It is not improbable that the various other elements were also present as chlorides. The recent work of Head,[15] Buerger,[16] and Newhouse [17] has shown that cavities in galena and sphalerite from Joplin contain a concentrated solution of NaCl and some $CaCl_2$. As this solution was imprisoned in these minerals at the time of their formation, its contents indicate something as to the composition of the solution that was *adjacent to the growing crystal.* It is doubtful, however, whether the entire solution in the cavity had the high concentration of chlorides indicated by the studies of Buerger and Newhouse.

The H_2S and CO_2 (if present) acted as dispersive agents, preventing reactions with the metals until the solutions reached the open ground, whereupon the change in pressure permitted the liberation of the gases. Rapid deposition of silica followed, and the H_2S could react with the lead and zinc chlorides to form the sulphides. The liberated chlorine could have united with any sodium present to form NaCl. The sodium might also have come in as Na_2S, but as this compound is a sulphide of a strong base and a weak acid, it is highly hydrolized and probably does not occur in such solutions. Zies [18] has shown that H_2S is a constituent of the fumarole gases in the Valley of Ten Thousand Smokes, and that it is more stable at the lower temperatures of

[15] Head, R. E.: The Cleavage Surfaces of Galena. Amer. Min., vol. 16, p. 345, 1931.

[16] Buerger, M. J.: The Negative Crystal Cavities of Certain Galena and their Brine Content. Amer. Min., vol. 17, p. 228, 1932.

[17] Newhouse, W. H.: The Composition of Vein Solutions as Shown by Liquid Inclusions in Minerals. Econ. Geol., vol. 27, p. 419, 1932.

[18] Zies, E. G.: The Fumarolic Incrustations and their Bearing on Ore Deposition. Nat. Geog. Soc. Tech. Papers, Katmai Series, vol. 1, No. 4, pp. 4, 12, 1929.

300°–100° C. He found, furthermore, that a fumarole which formerly deposited high-temperature magnetite from an acid solution *deposited a siliceous crust containing well developed crystals of galena, sphalerite, and copper sulphides* as the temperature decreased to 300°–100° C. It is more than significant that similar materials (silica, sphalerite, and some galena) represent the first materials deposited in the Tri-State area, and that the temperatures were within the same range as those in the fumaroles of Alaska. Furthermore, it is much more probable that the sulphur in the solutions in the Tri-State district was there as H_2S (rather than Na_2S), which reacted with the $PbCl_2$ or $ZnCl_2$ to form PbS, ZnS, and HCl. Any sodium originally present would unite with this Cl to form NaCl, or with calcium to form $CaCl_2$.

As the dominantly sulphide depositional phase passed, the carbonate, dolomite, was formed. It is probable that the incoming solutions contained Mg (as the chloride or carbonate) and that this Mg reacted with the $CaCO_3$ of the limestone to form dolomite. The $CaCO_3$ liberated (there was no increase in the volume of the wall rock) was reprecipitated as calcite or dolomite. The last phase of quartz deposition accompanied or followed the deposition of the carbonates, and the deposition of marcasite, pyrite, chalcopyrite, and a last generation of sphalerite closed the mineralization.

Mineralization occurred dominantly in the more open spaces within the breccias, but some replacement of limestone occurred. The solutions filled the open spaces, for the most part, but could not penetrate far into the Pennsylvanian shales, although they did so to a limited extent.

The essentially common level occupied by the major ore bodies throughout the district resulted as the solutions spread along the level of the solution channels. Where broken zones extended above or below this level, the ore solutions followed them.

SUMMARY.

The writer agrees with Weidman and Fowler and Lyden that the zinc-lead ores of the Tri-State area are of magmatic origin.

He does not agree with them in thinking that the great abundance of chert of the Boone formation is secondary in origin, but holds that it is primary or syngenetic and that it was deposited chemically on the sea floor at the time of the deposition of the enclosing limestone beds. This view is in keeping with the latest theories as to the origin of chert, and also does not require explanations that have little evidence to support them and that are entirely out of keeping with the sedimentary history of the Tri-State district and of the Burlington (Boone) formation elsewhere.

The brecciated areas are due to the limestone in the upper part of the Boone formation having been dissolved by ground water and the insoluble and broken chert left behind, rather than to mechanical forces acting along shear zones or to folding, as postulated by Fowler and Lyden and Weidman. Their explanations of the cause of these mechanical forces are far from convincing. Furthermore, the origin of the brecciated areas by underground water explains the sinuous course of the "runs" (they were water channels), the "circles" with their Pennsylvanian deposits, and the vertical zones, whereas these courses are not satisfactorily accounted for by any hypothesis employing mechanical forces.

After the breccias were formed, magmatic solutions deposited the jasperoid and accompanying sulphides and later minerals in the porous brecciated zones. The solutions were prevented from passing upward by the overlying Pennsylvanian shales and so spread out laterally along the nearly horizontal brecciated zones.

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THE COMPOSITION AND ORIGIN OF CERTAIN COMMERCIAL CLAYS OF NORTHERN IDAHO.[1]

EDWARD L. TULLIS AND F. B. LANEY.

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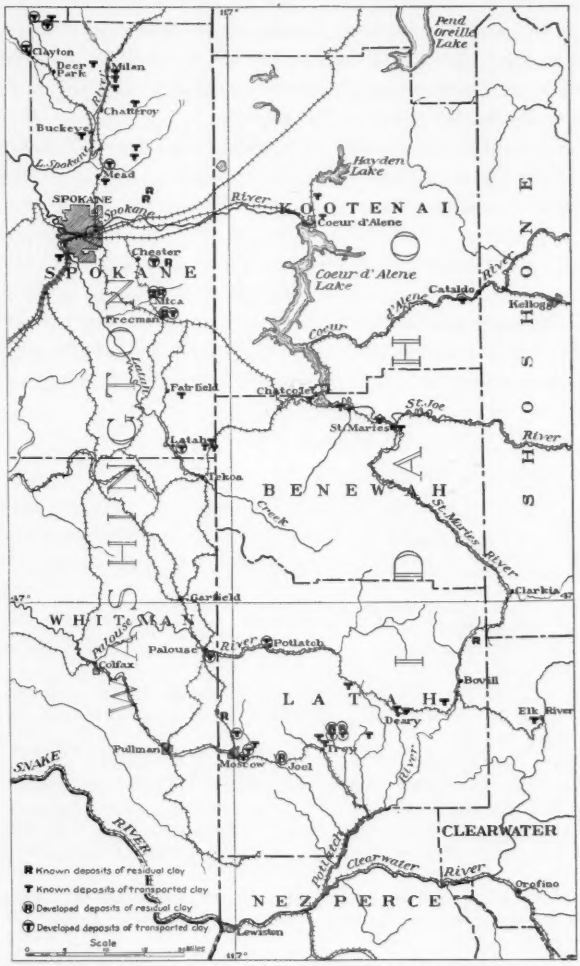
INTRODUCTION.

THERE has been an increasing scientific and commercial interest in the extensive residual and transported white clay deposits of northern Idaho and eastern Washington, at least partly because they constitute the most important non-metallic mineral resource of the region. Previous work has been either of a technological and commercial nature, or else, if of scientific interest, has been limited in scope. For these reasons a further study of the clays was thought desirable.

[1] Published by permission of Dr. John W. Finch, Director of the Idaho Bureau of Mines and Geology, to whom the writers are indebted for opportunities for field studies and the use of equipment.



WHITE CLAY DEPOSITS OF NORTHERN IDAHO AND EASTERN WASHINGTON

FIG. 1.

As may be seen from Fig. 1, white clays are found in Latah, Kootenai, and Benewah counties, Idaho, and in Spokane and Whitman counties, Washington, named in the order of apparent importance in their respective states. A study of those in Latah County forms the basis for this paper.

Previous Work.—Shedd [2] first studied the clays of Washington and Soper [3] later called attention to the adjacent deposits in Idaho. Skeels and Wilson [4] have given valuable data on the occurrence and properties of the Idaho clays. A general discussion of the deposits of both states was given by Wilson, Bennett, and Heath,[5] and at about the same time an important publication by Wilson [6] gave the firing properties and possible uses of the clays of Washington. The mineralogy and origin of certain of the deposits have been studied by Goodspeed and Weymouth [7] and by Hite.[8]

GENERAL GEOLOGY.

The oldest rocks exposed in the area are schists, gneisses, and quartzites of the Belt series of Algonkian age. These have been intruded by an extended arm of the Idaho batholith believed to be of late Cretaceous or early Eocene age.[9] However, it has been

[2] Shedd, Solon: The Clays and Clay Industry of Washington, pp. 168–229. State College, Pullman, Wash., 1910.

[3] Soper, E. K.: Fire Clays in Northern Idaho. Jour. Amer. Ceram. Soc., vol. 1, pp. 94–98, 1918.

[4] Skeels, Frank H., and Wilson, Hewitt: A Preliminary Report on the Clays of Idaho. Idaho Bur. of Mines and Geol., Bull. 2, pp. 24–28, 31–58, 62–65. Moscow, Idaho, 1920.

[5] Wilson, Hewitt, Bennett, A. L., and Heath, F. T.: Preliminary Report on the Residual Kaolin and Feldspar in the Pacific Northwest. Jour. Amer. Ceram. Soc., vol. 6, no. 3, 475–490, March, 1923.

[6] Wilson, Hewitt: The Clays and Shales of Washington, Their Technology and Uses. Univ. of Washington, Engin. Exper. Sta. Ser., Bull. 18, pp. 11–28, 60–62, 67–72, 1923.

[7] Goodspeed, G. E., and Weymouth, A. A.: Mineral Constituents and Origin of a Certain Kaolin Deposit near Spokane, Washington. Jour. Amer. Ceram. Soc., vol. 11, no. 9, 687–695, Sept., 1928.

[8] Hite, Thomas Hugh: The Origin of Certain Clays of Latah County, Idaho. A thesis, University of Idaho, Moscow, Idaho, 36 pp., 1930.

[9] Umpleby, Joseph B.: Geology and Ore Deposits of Lemhi County, Idaho. U. S. Geol. Surv. Bull 528, p. 43, 1913.

suggested recently [10] that the time of intrusion was late Jurassic or early Cretaceous. In Latah County this intrusive rock, the parent mass of most of the white clays, forms the Thatuna Hills. By middle Miocene time the area had been reduced to a mature topography, remnants of which may be seen in the hills of granitic and metamorphic rock that protrude above the basalt flows that began to pour in at about that time. The lava filled the valleys and altered the drainage with the formation of many lakes and ponds.

Studies in the Spokane area have shown that, preceding the outpouring of the basalt, the flora contained forms characteristic "of deep, fertile, well-watered soils, humid atmospheric conditions, and equable summer temperatures." [11] Deep weathering is found in all the pre-Miocene rocks of the region, even on the steeper hill-sides. The present residual clay deposits are found on the more gentle slopes, either beneath the basalt or at no great elevation above that more resistant rock, which preserves a low gradient and protects the softer material from erosion. The transported clays from the original bodies are found slightly farther down the hills and lying either between or above the basalt flows.

The present surface is underlain by a thick mantle of wind-blown soil that has been deposited upon the basaltic plateau.

RESIDUAL CLAYS.

There are two types of residual clay in the region, one formed from granitic and metamorphic rocks, and the other from basalt. Conclusions as to mineral changes and origin of the first type are based upon studies of the deposits near Troy and near Joel.

Microscopic examination has shown [12] that the granitic rock contains the following minerals: quartz, plagioclases, orthoclase, microcline, microperthite, biotite, muscovite, and accessory min-

[10] Ross, C. P.: Mesozoic and Tertiary Granitic Rocks in Idaho. Jour. Geol., vol. 36, 673–693, 1928.

[11] Berry, Edward Wilbur: A Revision of the Latah Formation. U. S. Geol. Surv. Prof. Paper 154, p. 234, 1929.

[12] Hite, Thomas H.: *Op. cit.*, pp. 11, 18.

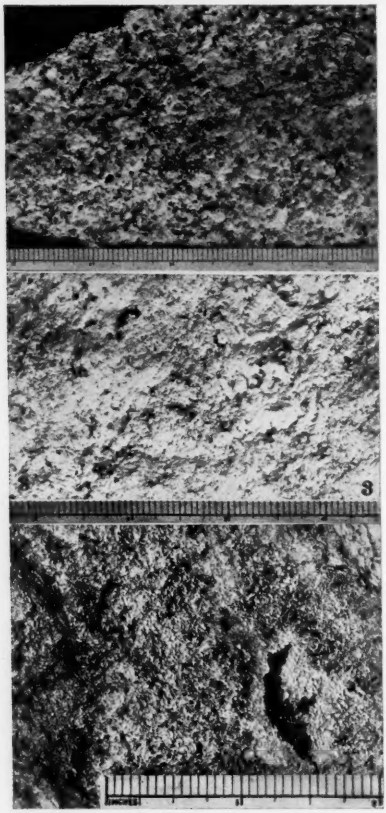


FIG. 2. Hand specimen of fresh rock from which the clay has been formed. Dark patches are biotite.

FIG. 3. Hand specimen of clay formed from the granitic rock, with original texture preserved. The biotite has been completely bleached. Darker areas are quartz.

FIG. 4. Hand specimen showing crystals of kaolinite formed from metamorphic rock. The large black area (lower right) is stained with manganese oxide. The smaller dark areas are colored by iron oxide.

erals such as apatite, zircon, black ores, xenotime, and monazite. A zeolite was noted in one thin section of slightly weathered aplite. A hand specimen of unaltered rock from near Troy is shown in Fig. 2. The original metamorphic rocks were a gneiss containing a considerable amount of fine to medium-sized quartz grains, and a schist composed chiefly of biotite, with some fine quartz in places. The high degree of decomposition prevents a more complete determination of the constituents.

The second type of residual clay, that from basalt, appears to have been formed in a different manner than have the granitic clays, and will receive further study.

Megascopic Appearance of the Clay.—The texture of the clay formed from the granitic rock is preserved, as may be seen in Fig. 3. The color varies from white to tan, caused by iron carried in surface waters descending from the covering of basalt which overlies portions of the deposits.

The clay formed from the metamorphic rocks has a color of mottled and banded cream, white and pink, if derived from the gneiss, and of purple and red, if from the biotite schist. The lighter colored material generally consists of rather fine quartz grains set in a matrix of clay, giving a granular character. Most of the darker clay is decidedly micaceous in appearance.

One of the most interesting features of the clay formed from the gneiss and schist is the presence of kaolinite crystals in bands that extend vertically through the deposit. The crystals (Fig. 4) vary in diameter from microscopic size up to 1½ mm. and have been noted up to 2 cm. in length, with the elongation roughly perpendicular to the trend of the bands in which they occur.

Decomposition of the Minerals.

The clay minerals are formed from feldspar, biotite, and muscovite. The final mineral produced is kaolinite but intermediate products were observed in the case of each of the three mineral sources. Feldspar may decompose either into a sericitic mineral or into an isotropic kaolinitic material. This isotropic substance may also form from the sericitic mineral. Either may

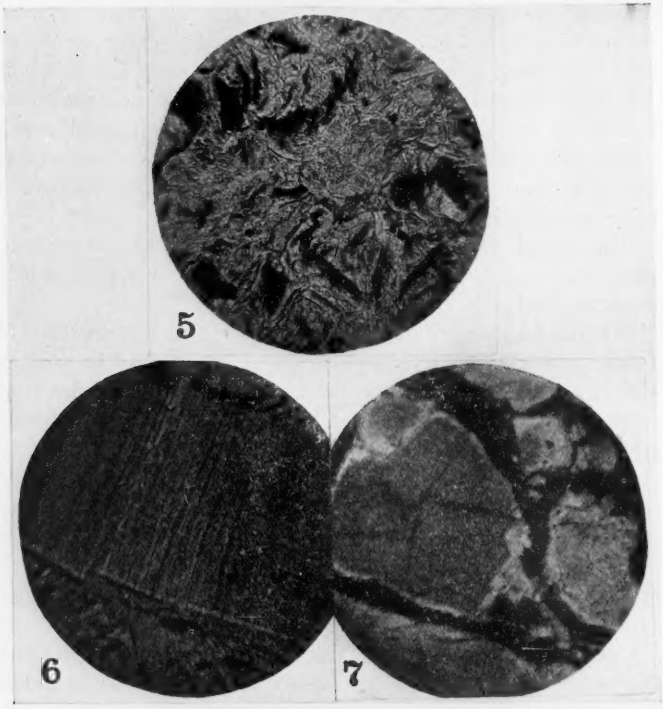


FIG. 5. Photomicrograph of thin section showing biotite changing to kaolinite. The biotite is pleochroic. Plane polarized light. × 38.

FIG. 6. A portion of a kaolinite crystal formed from biotite. × 165.

FIG. 7. Photomicrograph of thin section showing quartz replaced by kaolinitic material. Crossed nicols. The isotropic substance is clay material. All the pieces of quartz are in optical orientation and are parts of one crystal. × 100.

change directly into kaolinite. Muscovite forms kaolinite through stages of the sericitic mineral as in the case of feldspar. Biotite either changes to kaolinite directly, with a gradual change in physical and optical properties, or it may pass into an isotropic substance which in turn may form kaolinite. Detailed descriptions of the changes are given below.

Decomposition of Biotite.—The change from biotite to kaolinite will be discussed first because this has produced material best suited for study. As shown in Fig. 5, the largest kaolinite crystals appear to have formed in all cases directly from biotite without any intermediate products. The kaolinite crystals in Fig. 4 and 6 also have this origin. The original black biotite turns golden-brown in color, with almost complete loss of elasticity. At the same time there is a gradual decrease in the β and γ indexes of refraction and an accompanying loss of pleochroism, but most of the biotite retains part of its pleochroism until it is completely altered.

The end product, kaolinite, has good basal cleavage and also parting parallel to the crystallographic c-axis. The crystal faces parallel to this direction were not identified. The basal cleavage faces have a pearly luster and the side faces a waxy luster, and the color varies from white through cream to a pinkish-tan. The chemical composition of the larger hand-picked crystals is given in analysis 3 of Table I. Their specific gravity is 2.57.

In some cases there seems to be a tendency for the kaolinite crystals to develop a radiating arrangement with their c-axes arranged perpendicularly to the c-axis of the biotite. The kaolinite is biaxial and negative with an optic angle variable from small to large. The acute bisectrix is approximately normal to the basal cleavage and the maximum extinction angle measured against this cleavage is 3°. The indexes of refraction are $\alpha = 1.561$, β and $\gamma = 1.565 \pm 0.001$. These data agree with the optical properties of kaolinite as given by Ross and Kerr.[13]

Kaolinite forms from biotite also through intermediate stages. One such stage is a green mineral that resembles chlorite in the hand specimen but does not have the optical properties of chlorite. It in turn may pass into fine kaolinite crystals apparently not exceeding microscopic size. The mineral is green in thin section, is non-pleochroic, and is biaxial and negative, with an optic angle smaller than that found for most kaolinite crystals. Its β and γ indexes are 1.571 ± 0.005.

[13] Ross, C. S., and Kerr, P. F.: The Kaolin Minerals. U. S. Geol. Surv. Prof. Paper 165–E., p. 168, 1931.

TABLE I.

CHEMICAL ANALYSES OF CLAY MINERALS FOUND IN THE RESIDUAL CLAYS OF LATAH COUNTY, IDAHO.

R. V. Lundquist, Analyst.

Sample No.	1	2	3	4	5	6
Moisture below 100° C.	0.49	*	*	*	*	
Ignition loss	5.0	14.4	14.2	12.6	14.2	14.0
SiO_2	60.6	49.4	47.3	41.1	48.1	46.5
Al_2O_3	23.8	35.3	37.4	31.5	37.7	39.5
Fe_2O_3	0.3	1.14	1.55	13.9	0.71	
MnO	*	*	*	0.14	*	
CaO	0.19	0.19	0.28	0.09	0.46	
MgO	0.22	0.07	0.22	0.14	0.35	
Na_2O	0.32	0.00	0.32	0.06	0.18	
K_2O	8.77	0.27	0.00	0.06	0.08	
Total	99.8	100.8	101.3	99.6	101.8	100.0
$SiO_2:R_2O_3$		233:100	209:100		214:100	200:100

* Not determined.

1. Sample of partially decomposed crystals of orthoclase, from altered pegmatite dike in clay deposit near Joel, Idaho.
2. Entirely decomposed feldspar, from dike in clay deposit near Joel. The material was hand-picked and checked for visible impurities under the binocular microscope. The excessive SiO_2 over that required for kaolinite is perhaps accounted for by the presence of colloidal SiO_2.
3. Kaolinite from Joel; hand-picked under the binocular microscope. Formed from biotite.
4. Partially decomposed biotite, from highly altered biotite schist near Joel. Hand-picked under the binocular microscope.
5. Kaolinite from Joel; hand-picked under the binocular microscope. Formed from biotite.
6. Theoretical composition of kaolinite.

A type of decomposition of biotite not observed in the granitic rock was noted in the biotite schist. The mineral formed resembles biotite that has been stained red, but all the flakes have expanded and the folia are spread apart. The pearly luster on the basal cleavage and the waxy luster on the side faces cause the substance to have the appearance of kaolinite. Analysis 4, Table I, gives its chemical composition. The specific gravity is 2.79, a figure between the values for biotite and kaolinite. This material changes to kaolinite with a chemical composition given by

analysis 5, Table I. All these determinations were made on hand-picked crystals. The substance is isotropic and has a variable index of refraction ranging from 1.563 to 1.579. Scattered through the isotropic material and also more sparingly through the kaolinite crystals are numerous globules of an anisotropic substance, with a more intense red color than that of the isotropic or of the kaolinitic background. The globules, which are probably concentrations of iron oxide, have an index of refraction of 1.592 ± 0.002.

Decomposition of Feldspars.—The material derived from the feldspars shows varying degrees of decomposition from hard, friable crystals preserving the cleavage of the original mineral and having a chemical composition as given under sample 1, Table I, to a soft, entirely decomposed substance, the composition of which is shown in sample 2, Table I. The first analysis, made on portions of large phenocrysts, is similar to those given by Bayley [14] for samples of semi-kaolinized orthoclase from North Carolina. The second analysis was made upon fine-grained white clay picked by hand from entirely decomposed feldspar crystals. No attempt was made to further refine this hand-picked material, except to remove impurities visible under the binocular microscope. The silica in excess of that required for kaolinite is perhaps explained by the presence of colloidal SiO_2. The specific gravity of this clay was determined to be 2.53.

The partially decomposed feldspar preserves the optical properties of the fresh mineral, but there is an abundant development of both isotropic and sericitic material along the cleavage cracks. The sericitic mineral probably represents the first stage in the change of the feldspar and the next stage is the formation of fine white clay, which appears isotropic under the microscope. The evidence for this sequence is not conclusive, but instead the isotropic and sericitic minerals may represent parallel alterations, because both were observed changing directly to kaolinite, the sericitic material more sparingly.

The mineral resembling sericite is probably the same as the

[14] Bayley, W. S.: The Kaolins of North Carolina. No. Car. Geol. and Econ. Surv., Bull. 29, p. 18, 1925.

"muscovite-like kaolin mineral" described by Ross and Kerr,[15] but it is believed that perhaps two such minerals occur with different indexes of refraction. A number of cleavage flakes were determined to have β and γ indexes of 1.580 ± 0.001, while for many other pieces these indexes were found to be 1.571 ± 0.005. In all cases the mineral was determined to be biaxial and negative with a small optic angle. The index 1.571 ± 0.005 corresponds with the values given by Ross and Kerr, but these men determined $2V = 60° \pm 5°$, which is an angle considerably larger than that in the case of most of the material studied. It would appear that the two sets of indexes are characteristic of two definite steps in the alteration of the feldspar, but it is possible that there is only a uniform change in the value of the index of refraction. The "muscovite-like mineral" with the lower index alters generally to the isotropic material mentioned above, but in a few instances was found to form crystalline kaolinite directly, with the crystallographic c-axes of the two minerals perpendicular to each other.

However, the larger percentage of the kaolinite crystallizes, not from the sericitic mineral, but from the isotropic material that has already been mentioned as a decomposition product of the sericitic mineral. This isotropic substance predominates in most of the clay, except in the completely altered portions of the deposits where the whole mass has changed to kaolinite in the form of crystals that do not exceed microscopic size. The mean indexes of refraction of both the isotropic material and the fine crystalline kaolinite are in the range 1.557 ± 0.004, but of the two minerals, the kaolinite always has a slightly higher index. The naming of the isotropic substance is a problem. Its index of refraction is that either of halloysite or of kaolinite composed of particles of colloidal size. In fact the distinction between such kaolinite and halloysite is not clear.[16]

Decomposition of Muscovite.—In only a few cases was muscovite observed to be completely changed to kaolinite. On the other hand, it was found to retain its original optical properties

[15] *Op. cit.*, pp. 172–173.

[16] Ross, C. S., and Kerr, P. F.: *op. cit.*, p. 154.

only in the center of the flakes from the deeper parts of the deposits. The freshest muscovite to be found in the clay has β and γ indexes of 1.592 ± 0.005. In each flake there is an outward gradation into sericitic material having the same optical properties as the "muscovite-like mineral" described as an alteration product of feldspar. Beyond this stage the alteration appears to be exactly as in the case of the feldspars.

Replacement of Quartz.—Quartz is in some instances replaced by kaolinitic material as shown in Fig. 7.

Origin of the Residual Clays.

Two major theories of the formation of the residual clays have been advanced. One states that they were formed by hydrothermal solutions given off perhaps as emanations of the intrusion that is the parent rock of the granitic clays. The other states that the clays are a product of weathering. The writers believe this theory to be the true one. The climate preceding the outpouring of the basalt in Upper Miocene time was such as to have caused the deep decomposition observed in the present clay deposits. The points of evidence supporting the theory of weathering follow:

1. One mineral found universally in the clays is kaolinite, which is commonly believed to be formed in abundance only by processes of weathering. In fact, Ross and Kerr recently have stated that instead of kaolinite, other closely related, but distinguishable, minerals would be formed by hydrothermal processes.[17]

2. The clay contains no accessory minerals commonly associated with hydrothermal alteration that are not also found as abundantly in the original unaltered rock.

3. No channels along which hydrothermal solutions would have passed are present.

4. The decomposition is uniform over large areas, up to a square mile in one instance. Extensive and deep weathering is characteristic of all the older rocks of the area.

[17] *Op. cit.*, pp. 174–176.

5. There is no increase of alteration with depth and there appears to be an actual decrease. In the deeper parts of the abandoned pit near Troy the biotite is less altered than in the upper portions, and the centers of the muscovite flakes are fresher at depth. The change from isotropic clay material, of the former feldspar, to crystalline kaolinite is less complete in the deeper parts of the deposit.

A combination of the two theories of origin suggests that the clay has been formed by weathering in restricted areas where previous hydrothermal action had caused partial alteration and rendered later weathering more effective. The surrounding area and the more elevated regions were studied with this idea in view, but no supporting evidence was found. The absence of clay around much of the border of the older rocks has been caused by erosion. No residual deposits are found along the bottoms or slopes of the valleys but only at the upper margins of the basaltic pleateau where the clay has been preserved from removal.

Goodspeed and Weymouth [18] have offered evidence to show that the kaolinite in the clay at Freeman, Washington, a deposit somewhat similar to that at Joel, is of primary origin. The writers have not studied the deposit at Freeman, but the suggestion certainly does not apply to the deposits of Latah County. Here the kaolinite has been formed by progressive changes as outlined in preceding paragraphs.

TRANSPORTED CLAYS.

So far as determined, the mineralogy of the transported clays is the same as that of the residual bodies from which they were derived. The possibility of the formation of other minerals by further change is to be investigated in later studies. Some layers within the clay are sandy, others contain considerable mica, and still others may be composed of almost pure clay substance as shown by the chemical analysis of Sample 1, Table II. Microscopic examination of this specimen showed the chief constituent to be isotropic clay material containing a few small scattered crystals of kaolinite.

[18] *Op. cit.*, p. 694.

TABLE II.

CHEMICAL ANALYSES OF CERTAIN CLAYS OF LATAH COUNTY, IDAHO.

R. V. Lundquist, Analyst.

Sample No.	1	2	3	4	5	6	7	8	9	10
Moisture below 100° C.	2.30	1.71	0.78	0.57	0.45	0.79	1.07	1.19	1.07	1.19
Ignition loss	14.5	13.2	8.7	7.2	7.4	6.0	7.7	7.3	9.9	9.9
SiO_2	46.2	49.5	66.5	68.3	71.1	72.8	69.1	71.9	62.3	61.8
Al_2O_3	35.7	34.0	22.2	20.5	20.0	19.5	20.0	19.7	26.1	26.2
Fe_2O_3	0.1	0.4	1.4	1.9	0.6	0.71	1.9	0.29	1.11	1.13
MnO	*	*	*	*	*	0.00	*	*	*	*
CaO	0.37	0.19	0.45	0.38	0.28	0.09	0.38	0.19	0.27	0.28
MgO	0.21	0.22	0.21	0.36	0.22	0.07	0.07	0.14	0.21	0.22
Na_2O	0.67	0.44	0.26	0.32	0.37	0.03	0.05	0.00	0.11	0.21
K_2O	0.03	0.60	1.15	1.59	0.58	1.28	0.17	1.01	0.30	0.60
Total	100.1	100.3	101.7	101.1	101.0	101.3	101.4	101.7	101.4	101.5

* Not determined.

1. Very pure transported clay from near Troy, Idaho.
2. Residual granitic clay washed through a 200-mesh screen. From near Troy.
3. Residual granitic white clay from the vicinity of the Idaho Fire Brick Company's new pit near Troy.
4. Residual granitic clay colored tan by iron oxide. From abandoned pit of the Idaho Fire Brick Company one mile to the east of Sample 3.
5. Residual granitic white clay from a point located between the places at which Samples 3 and 4 were taken.
6. Residual white clay from the decomposed aplite and pegmatite dikes near Joel, Idaho.
7. Residual white clay formed from the decomposition of a gneiss at Joel, Idaho.
8. Transported white clay from an abandoned pit of the Idaho Fire Brick Company, near Troy.
9. Transported clay from the vicinity of the pit of the Moscow Fire Brick and Clay Products Company, a mile east of Moscow, Idaho.
10. Transported clay from pit two miles north of Moscow.

Note: These are not intended to represent average analyses of the deposits, but were made on specimens that were taken as representative samples for mineralogical study.

Origin of the Transported Clays.—Some of the beds of transported clay are found between basalt flows and others are found above the flows. It is clear that these deposits were formed in

lakes and ponds caused by the damming of the streams with basalt in Miocene time.

ECONOMIC FEATURES.

Reserves.—It would appear from the accompanying map that the clays underlie larger areas in Washington than in Idaho, but this is probably only apparent because the known deposits are most numerous in the more thickly settled and better developed regions. Any statements as to the extent and development of the clays in eastern Washington would be only a repetition of the discussions of others.

Definite estimates of reserves for the deposits in Latah County are impossible, but some idea may be gained by a statement of their areal extent. Even this factor is not exactly known because extensions of the bodies are occasionally being discovered in new road cuts, excavations for buildings, etc. Near Deary the clay underlies an area of from 40 to 100 acres. Two of the deposits are exposed to a depth of 20 feet in railroad cuts and another is reported to have been drilled to a depth of 100 feet. The transported deposits near Troy are more numerous but perhaps not so extensive as those near Deary. From five to six miles northeast of Troy a residual deposit is known to underlie almost a square mile and has been drilled to depths of 60 and 70 feet at two widely separated points, without the bottom of the deposit being revealed in either case. The residual clay near Joel underlies an area of about 40 acres and has been penetrated to a depth of 40 feet. Near Moscow, transported clay outcrops in a number of places from two to three miles apart, and although the deposits are probably not continuous under the entire intervening area, their total extent must be more than 100 acres. The depth in one case is known to exceed 35 feet.

Development.—Clay has not been removed from the deposits near Deary and near Potlatch for a number of years. The pits in the transported clays near Troy have been idle for two or three years but the adjacent residual deposit is under active development. The pit in the residual clay at Joel is only temporarily idle. Near Moscow, work was carried on at three different places during the summer of 1932.

Utilization.—The white clays, both residual and transported, of northern Idaho and eastern Washington are used in making the following products: fire brick of all types, face and common brick, fancy enameled brick, heavy structural tile, roofing tile, drain tile, sewer pipe, terra cotta, flower pots, and stoneware of various kinds. The possible uses are even more varied than these, particularly if the clays are refined by the removal of quartz and mica, both of which could be used as by-products. Wilson [19] states that the clays of this region, together with those of western Washington, are suited to be the source of a large part of the Pacific Northwest supply of porcelain and whiteware as well as the stoneware and pottery, including bath, laundry and sanitary wares, enamelled metal wares and cheaper household electrical insulators. The market of which the clays are deserving is much greater than they find at present, including at least all the country west of the Rocky Mountains.

Chemical analyses of specimens from both residual and transported deposits in Latah County are given in Table II.

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[19] *Op. cit.*, pp. 19–20.

A CHALCOCITE DEPOSIT, GREAT BEAR LAKE, CANADA.

CHRISTOPHER RILEY.

INTRODUCTION.

THE occurrence of native copper on the Arctic coast of the North West Territories of Canada has been known since Hearne's explorations in 1780. Since that time there have been occasional investigations of a sketchy nature. The advent of the airplane in northern prospecting, however, has made these regions comparatively accessible, and in the past few years active prospecting has been carried on. This has resulted in the discovery of several interesting copper deposits near Dismal Lakes in the Coppermine River region and on Great Bear Lake to the southwest.[1]

So far, it seems that native copper is less likely to be found in commercial amounts than deposits of bornite, chalcocite, and chalcopyrite. Several promising occurrences of these minerals, in veins associated with quartz and carbonates, have been located, but no important finds of native copper have yet been reported. Unfortunately, the price of copper dropped soon after the discoveries were made and little information is as yet available as to the extent of mineralization. The small amount of work that has been done indicates that this area may become of importance when higher copper prices obtain.

In the summer of 1932 the writer examined a copper deposit lying south of Hunter Bay on Great Bear Lake, which though small, seems unique enough to merit description. Further, its

[1]Gilbert, G.: Copper on the Coppermine River. ECON. GEOL., vol. 26, p. 96. Norrie, J. P.: Prospecting in the Great Bear Lake-Coppermine River Area. Can. Min. and Met. Bull. no. 227, p. 349, 1931. Duncan, G. G.: Exploration in the Coppermine River Area. Can. Min. and Met. Bull. no. 227, p. 363, 1931. Kidd, D. F.: Great Bear Lake-Coppermine River Area. Canada Geol. Surv. Summ. Rept., Pt. C, p. 60C, 1931.

occurrence may shed some light on the origin of the deposits of Hunter Bay and even on those of the Coppermine River area.

This deposit lies on Hogarth Island on the east shore of McTavish Arm of Great Bear Lake about 15 miles southwest of the Hunter Bay copper claims. The island is divided almost in two by a deep bay entering it from the southeast. It is in this bay, on its southwest shore, that the copper occurs.

FIELD RELATIONSHIPS.

The deposit consists of a vein of what appears to be massive chalcocite, with a little bornite, in sharp contact with granite country rock. The vein branches off at right angles from the south side of a well defined fault with a north-south strike and vertical dip. This fault is clearly defined for at least one-quarter of a mile. The strike of the vein is N. 70° E. and its dip is vertical. The width of the chalcocite varies from one to two feet and it is exposed over a distance of one hundred feet along the strike. The contact with the granite is quite sharp and regular except where small offshoots branch into the granite for short distances from either side of the vein.

The vein material consists mainly of dense, massive chalcocite, with a little bornite and covellite visible. A few small stringers of quartz and barite form the gangue. The granite has been altered to a distance of four feet on either side of the vein. Fifteen feet from the vein there is a small lens of pyrite in the granite. Quartz occurs in considerable amounts in the fault from which the vein branches. A diabase dike lies about 100 feet north of the vein and has approximately the same strike. It is intersected by the main fault and is thus earlier than the mineralization.

MICROSCOPIC RELATIONSHIPS.

In polished section, chalcocite is seen to be the predominating mineral. Bornite forms about 20 per cent. of the mass, klaprotholite ($3Cu_2S.2Bi_2S_3$) is present in small, scattered grains, and covellite occurs along fractures and near the exposed sur-

faces in small amounts. The gangue consists of quartz, hematite-bearing quartz and barite.

Chalcocite and bornite in part show mutual crystal boundaries and, in a few places, a graphic intergrowth, but most of the chalcocite is the result of the replacement of bornite. There are several indications of this replacement. Small stringers of chalcocite traverse the bornite. Bornite and quartz which once existed in juxtaposition, are now separated by narrow, intervening bands of chalcocite that have replaced the bornite along these lines of easy permeability. Bornite commonly occurs as fragmentary, unreplaced remnants with ragged edges, in the chalcocite.

The rather rare mineral klaprotholite was identified by the usual etch and microchemical tests. Of these the KCN etch test was inconclusive and the color of the mineral has more of a bronze cast than that ascribed to most klaprotholite. This mineral shows mutual boundary relationships with the bornite and with some of the chalcocite. It is not replaced by either mineral. Covellite replaces both the chalcocite and bornite and is clearly of secondary origin.

Quartz appears to have formed throughout the whole period of mineralization, appearing as idiomorphic grains, also in mutual boundary relationships and as stringers cutting the metallic minerals. Hematite-bearing quartz is much less common than the clear quartz and occurs in stringers that cut the other minerals. Barite is also found as stringers of later origin. The approximate proportion of minerals and gangue is:—chalcocite 57 per cent., bornite 20 per cent., covellite 1 per cent., klaprotholite 2 per cent., quartz 20 per cent., and barite 1 per cent. The paragenesis is as follows:

The granite consists of quartz, orthoclase, oligoclase, biotite and accessory magnetite. Even in the least altered phases there has been some alteration of the feldspars. The orthoclase is somewhat clouded with kaolin and the plagioclase is much more altered and is so filled with hematite dust as to give it a distinctly brick-red color sharply contrasting with the gray orthoclase. This hematite appears to have been derived largely from the alteration of the biotite, which is nowhere fresh and is ordinarily surrounded by a halo of hematite dust derived from its alteration.

Two feet from the chalcocite vein the feldspars of the granite are totally altered to white mica and kaolin and the biotite is bleached nearly white. The granular texture remains. Adjacent to the vein the rock has lost all indication of its original texture and nothing remains but kaolin, hydrous iron oxide and quartz, some of which, occurring in tiny stringers, appears to have been introduced. At least part of the original iron content of the rock still remains as limonite and thus the bleaching of the rock that is found in some deposits of rich copper minerals, particularly the deposits of native copper, has not taken place here.

RELATIONSHIP TO OTHER DEPOSITS OF THE AREA.

There is evidence that the Hogarth Island deposit is related in origin to other copper occurrences of Great Bear Lake and perhaps also to those of the Coppermine River area. There are many small copper deposits on the islands and mainland in the vicinity of Hogarth Island, in which the chief mineral is chalcocite associated with quartz. This mineral appears either as stringers or as blebs in quartz veins. A larger deposit of this type is found at Hunter Bay, where predominating chalcocite and bornite with smaller amounts of chalcopyrite and famatinite are associated with a huge quartz vein some 500 feet wide.[2] This quartz vein lies in part in granite and in part in volcanic rocks. Again, the Hogarth Island deposit is similar to those of the Coppermine area in that the latter consist mainly of chalcocite and bornite. There is also some chalcopyrite, with small amounts of quartz and a

[2] Kidd, D. F.: *Op. cit.*, p. 55C.

carbonate mineral. In a polished specimen from the Coppermine, chalcocite was seen replacing bornite as at Hogarth Island.

ORIGIN.

The evidence that is available favors a hypogene origin for the Hogarth Island deposit. This evidence may be summarized as follows:

1. Klaprotholite is doubtless a mineral of hypogene origin and its mutual boundary relationships to chalcocite and bornite suggest that these minerals have a similar and contemporaneous origin.

2. The veinlets of quartz, hematite-bearing quartz, and barite that cut the chalcocite veins appear primary and may represent a later phase of the mineralizing process that deposited the chalcocite. No secondary quartz was recognized in the district and hematite-bearing quartz of primary origin is common.

3. With the exception of a thin coating of malachite and a little covellite there are no suggestions of secondary action in the deposit.

4. The quartz veins of the nearby localities carrying chalcocite are primary and the chalcocite occurring in them as blebs probably came in with the quartz.

Considerable study has been made of copper deposits low in sulphur and iron.[3] Particularly in connection with the native copper deposits has it been suggested that the deposition of the native copper has been due to the oxidation of the mineralizing solutions by interaction with a host rock high in ferric oxide or with minerals in the host rock high in ferric oxide. It can here be argued that a somewhat analogous situation exists in the Great Bear Lake area. One of the most notable features of that district is a widely extensive hematite mineralization. This mineral occurs in various forms: as numerous tiny stringers and in veins up to eight inches wide; in considerable amounts injected

[3] Singewald, J. T., Jr., and Berry, E. W.: Johns Hopkins University Studies in Geology, no. 1, 1922. Bateman, A. M.: ECON. GEOL., vol. 18, pp. 122–166, 1923. Butler, B. S.: U. S. Geol. Surv. Prof. Paper 144, 1929. Bastin, E. S.: The Chalcocite and Native Copper Types of Copper Ore Deposits. ECON. GEOL., vol. 28, pp. 407–446 (in press).

into the pores of loose textured sediments; filling quartz crystals in vugs and in some of the quartz veins of the area, and also as coloring in some calcite veins. So far as could be ascertained, it is of a late generation compared with the other mineralization.

That hematite is closely associated with the deposition of the Hogarth Island chalcocite is evident from the presence of hematite in some of the later quartz stringers, and in this connection an interesting speculation might be made. Shortly after the formation of the vein commenced, resulting in the deposition of bornite, ferric oxide entered the solutions in amounts large enough to oxidize some of the sulphur, and, where pressure and temperature changes permitted, chalcocite was formed. The iron and the sulphur were carried off in the soluble form $FeSO_4$. Copper and sulphur decreased in amount before the ferric oxide did, so that the latter was deposited as hematite in quartz stringers late in the period of mineralization. As noted by Bateman [4] in the Bristol deposit, the presence of barite points to the reduction and solution of ferric oxide with the probable formation of ferrous iron and sulphates. Of these sulphates that of barium remained behind due to its relative insolubility.

The source of the solutions may well have been in the magma from which the widespread granite intrusions of the area emanated. The only intrusives of a different nature in the district are diabase dikes, and since the Hunter Bay copper deposits are closely related to the huge quartz veins there and since these veins are not likely to have been associated with the diabase, it seems reasonable to suppose that the copper minerals had their origin with the granite. Similar granite is found in the Coppermine and may have been the source of the chalcocite fissure deposits there. Where the solutions that formed the chalcocite penetrated the Coppermine River amygdaloidal flows, the sulphides were further reduced to native copper.

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[4] Bateman, A. M.: Primary Chalcocite, Bristol Copper Mine, Conn. ECON. GEOL., vol. 18, pp. 154 and 164, 1923.

EDITORIAL

WHAT IS ROCK?

Probably most economic geologists who have had occasion to act in a consulting capacity, in connection with engineering projects, have been appealed to by contractors or engineers to give an answer to the question which heads this editorial. The difficulty is one inherent in the use of a common English word in two senses—one that of familiar, and consequently more or less inexact, significance, and one with the meaning attached in the scientific nomenclature of geology.

In geology, any considerable mass of solid material which constitutes a part of the earth is classed as rock, irrespective of hardness. Thus Geikie states:

> The soft sand and mud of a modern lake-bottom differ in no essential respect from indurated ancient strata and may tell their geological story just as well. No line is to be drawn between what is popularly termed "rock" and the loose, as yet uncompacted débris out of which solid masses may eventually be formed. Hence, in a geological arrangement, the modern and the ancient, the loose and the compact, being one in structure and mode of formation, are all classed together under the common name of Rocks.[1]

Probably, however, nearly all geologists are sufficiently under the influence of popular usage to recognize some distinction between rock and residual soil, although most of them would acknowledge that the ice-cap of Greenland is rock.

In common language, *rock* ordinarily means a mass of solid, mineral material of considerable size and of some unspecified degree of hardness. Webster's dictionary, for example, gives as the first and general definition:—"A large concreted mass of stony material; a large fixed stone," etc. Ordinarily, *rock,* in

[1] Geikie, A.: Text Book of Geology, vol. I, pp. 159–160. 4th ed., 1903.

practical affairs, such as excavation, is the designation for material of such hardness that it can not be removed by pick and shovel without prior drilling and blasting. In popular usage, granite would unhesitatingly be called rock, although loose sand and gravel would not. Nevertheless, there are various intermediate stages, as between sand and sandstone, or between gravel and conglomerate, to which the term *rock* is not applicable with accuracy and unanimity. For example, in the Tertiary formations of California, certain sandstones are so soft that the non-geological person is at a loss whether to term them rock or sand. In such cases, the choice of the word to use may often depend upon more or less immaterial circumstances or predilections, as, for instance, the geological age of the deposit, or its relation to associated mineral materials.

It thus appears that the common significance of the word is far from definite. What may be "rock" to one man is not necessarily "rock" to another. The term is much too vague for use in classifying material for excavation, unless rigorously qualified. Even the use of the adjective "solid" does not overcome this objection, inasmuch as material may be very soft and yet undeniably solid.

It has frequently happened that where disagreement with respect to the class of material excavated has arisen between the contractors and those responsible for payments, a geologist has been called in, under the naïve supposition that he, if anyone, should know rock when he sees it. Engineers and contractors should realize, however, that the geologist's definition of rock is not what is required. The problem is ordinarily one that calls for an interpretation of loose or indefinite statements, and consequently demands the ability of a lawyer or of an expert in linguistic expression. All concerned generally know as much as is practically necessary about the character of the material, but they do not agree as to what the specifications mean.

A concrete example of the difficulty that may arise in classifying material to be excavated may be of interest. In connection with a minor contract on a very large engineering project, the following paragraphs are found in the specifications:

Rock Excavation.—All solid rock in place, which can not be removed until loosened by blasting, and all boulders or detached pieces of solid rock more than one-half cubic yard in volume. Solid rock under this class, as distinguished from soft or disintegrated rock under common excavation, is defined as sound rock of such hardness and texture that it can not be effectively loosened or broken down by hand drifting picks.

Common Excavation.—All material required to be excavated except solid rock, including earth, gravel, and also indurated material of all kinds, such as hardpan, cemented gravel, and soft or disintegrated rock, which may require blasting before removal by team-drawn scrapers or excavating machinery; also all boulders or pieces of rock, not exceeding one-half cubic yard in volume, that are detached or embedded in common excavation material.

Probably most geologists, in reading these specifications, will recognize their contradictory and confusing character, particularly the uncertainty introduced by the proviso that "common excavation" shall include "indurated material of all kinds," which "may require blasting." This proviso has the effect of including under "common excavation" material which might fairly be termed *rock* and which would otherwise reasonably be classed under "rock excavation."

As a matter of fact, the disagreement, which might have been expected, actually arose. The contractor was asked to accept payment at the low "common" rate for excavating material that could not be effectively removed by pick and shovel or by trenching machines, until it had been drilled and blasted. This material is a sedimentary breccia or fanglomerate, of probably Quaternary age, in which the rock fragments of various kinds range in size from sand grains to pieces up to 2 feet in length. The cementing material is probably not all of one kind. In some of the hardest and toughest varieties, it is plainly gypsum. Oxides of iron may also contribute in places to the general cohesion. Although gypsum is a soft mineral, it constitutes a tough and effective cement. Blasting of this material throws out masses over 6 feet in diameter. These blocks can be lifted by steam shovels into trucks and dumped, without falling to pieces, and the dumped blocks retain their angular form and in many instances show the drill-holes through which the blasting charges were introduced. By

at least nine persons out of ten, they would probably be termed "rock" in the popular sense, and they are certainly rock in a geological sense. In this particular contract, the classification as "common excavation" reduced the amount to be paid to the contractor by about $20,000.

What are the prerequisites against the occurrence of such an unfortunate situation? In the first place, it would appear that a geologist, who is likely to realize fully the difficulties inherent in the use of the work *rock*, should be consulted in the drawing of the specifications. Geologists, in general, should lose no opportunities of impressing upon engineers and contractors the necessity of greater care and accuracy in the drafting of specifications for excavation. More fundamentally, the attempt to found a distinction on the term *rock* should be abandoned. The essential thing is the relative cost of removing the material and the distinction should be based on the methods that can most effectively and most economically be employed. Material might be classed with reference to the necessity or non-necessity for blasting, the grades of explosives required, and the behavior with respect to standard equipment or tools used in excavating. If material must be drilled and blasted, the cost of removal is practically that of material commonly called "rock." It is of little practical consequence what such material is called, but it is of primary importance to classify it with reference to the cost of removal by standard methods efficiently employed. If a contractor can devise new and better methods, he is entitled to the reward resulting from his ingenuity.

F. L. RANSOME.

DISCUSSION AND COMMUNICATIONS

ORIGIN OF BAUXITE DEPOSITS.

Sir: Referring to the suggestion of C. H. Behre, Jr. in this JOURNAL[1] that bauxitization may occur where rocks are exposed to unusual amounts of humic or other acids, some circumstances brought to light during an investigation of the Tikhvin bauxites, U. S. S. R., seem to be partly corroboratory and partly contradictory to Professor Behre's suggestion.

The Tikhvin bauxites occur some three hundred kilometers east of Leningrad at several spots along a Devonian-Carboniferous unconformity. Undoubtedly, they were developed during the corresponding erosion period, both as a residual mantle on Devonian blue and green clays and as locally deposited sediments which may or may not have been derived from the residual bauxite. The residual bauxite grades into Devonian clay beneath. The associated Carboniferous rocks are a series of loose sands and clays about thirty meters in thickness, overlain by several hundred feet of Productus-bearing limestones. An occasional thin lens of bauxitic sediment exists in the sand-clay series, which series, along with the Devonian clay, is remarkable for being entirely unlithified, in spite of its age. Most of the bauxite is not of ore grade, being either clayey or lateritic.

The sedimentary nature of some of the thicker bauxite deposits was thought to be indicated by their being convex downwards, in lenses several hundred or a thousand meters in lateral extent, and by their having included seams of pitch black carbonaceous clay. The black clay lenses out at the margins of the bauxite basins and has initial dips reflecting the shapes of the basins. It seemed quite obvious that the basin-shaped bauxite deposits and the black clays are local deposits in swamps or ponds.

[1] Behre, C. H., Jr.: Origin of Bauxite Deposits. ECON. GEOL., vol. 27, pp. 678–680, 1932.

The black clay does not everywhere overlie the bauxite in the basins, but its occurrence in so far as discovered was restricted to the two such basins outlined. In many spots there is exact coincidence. The carbonaceous nature of the clay and the apparent conditions under which deposition occurred seem to be good evidence that humic acids were probably present.

However, except for the two examples just mentioned, there is no evidence to be drawn from most of the bauxite or its associated rocks that unusual amounts of humic or other acids were present during its development, as they do not contain notable amounts of carbonaceous material, sulphides or sulphates.

It seems as though there must be other important limiting factors to bauxitization besides the possible presence of humic or sulphuric acids; otherwise bauxite would be a common deposit in place of the fire clays so common under coal seams.

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REVIEWS

Principles of Historical Geology. By R. M. FIELD. Pp. 283, figs. 103; pls. 10 (in pocket). Princeton Univ. Press, 1933. Price, $3.50.

This book is unique among texts on historical geology in that for the first time it eliminates period-by-period chronological treatment and attempts to show the student how earth history has been deciphered by considerations of six districts, namely, the Grand Canyon, Niagara, Appalachians, Scottish Highlands, the Alps, and the Yellowstone Park and Big Horn Basin. Part II treats of these. Part I (80 pp.) is made up of:—The History of Stratigraphy; The Sedimentary Rocks; and The Technique of Correlation. The usual plates of dinosaurs and invertebrates are missing; five cuts only illustrate geologic life.

There is perhaps a place for such a book in elementary teaching but the reviewer wonders why it is called Principles of Historical Geology. The principles dealt with are few and the historical geology seems subsidiary; it might just as fittingly be called Principles of Structural Geology. For example, the Alps is one of the six districts chosen for study. Twenty-two pages are devoted to it; 5 text pages are introductory; 1½ are geographical; 10½ are descriptions of structure and stratigraphy; and 2½ historical geology.

As an elementary description of the geology of the six regions, the book is useful, although so much straight description makes rather dry reading. After the numerous errors in which it abounds are corrected, it might serve satisfactorily as collateral reading in a course in historical geology.

The preparation and make-up of the book, unfortunately, are unpleasing. The type is small and too closely leaded for easy reading and the paper receives the illustrations poorly. Halftones are absent and many of the line cuts are crudely drawn and most of them poorly reproduced; they were improperly prepared for the reductions used, and several variations of crude hand lettering give them a sloppy appearance; some diagrams are almost undecipherable. Even de Martonne's classical diagrams of anticlinal valleys (Fig. 73) have suffered in reproduction, as well as in credit to their source. Further carelessness is shown in the legend of Fig. 10, where shale and limestone (*rocks*) are listed as *sediments*. The legend of Fig. 33 is unintelligible. The plates of block diagrams (in pocket) are excellent.

Carelessness in text matter and proof reading is also noticeable; the first paragraph starts out with *origin* mis-spelled, and the same chapter ends (p. 20) with an out-of-date statement of the age of the rocks as determined by the radium method. Murchison instead of Sedgwick is credited (p. 9) for the Cambrian; Omar Khàyyam is stated (p. 10) to have described the retreat of the sea in the "second century," (this gives him a life of over one thousand years!); plateaus are bounded by "fixtures" (p. 81); Niagara (p. 116) flows "2,250,000 gallons per minute" (just 2.1 per cent. correct); Niagara Falls (p. 117) is "over 1500 feet less in height than Victoria Falls" (What a falls!); granite produces "aurioles"(!) of contact metamorphism (p. 178). Subaerial is consistently misspelled throughout, and correct dates are apparently inconsequential. To mention the errors noted* even in a cursory reading would more than fill this page.

ALAN BATEMAN.

Earth Oil. By GUSTAV EGLOFF. Pp. 158, figs. 41. Century of Progress Series, Williams & Wilkins, Baltimore, 1933. Price, $1.00.

This is another interesting volume added to the Science Progress series for popular reading. It gives a bird's eye view of the whole oil industry, including the antiquity of oil, geology, and origin, also locating, drilling, storage, transportation, and refining. A chapter on oil resources deals lightly with production figures and hints that no immediate shortage is in sight.

Stream Sculpture on the Atlantic Slope: A Study in the Evolution of Appalachian Rivers. By DOUGLAS JOHNSON. Pp. 142, figs. 21. Columbia University Press, 1931. Price, $3.25.

This volume, written by a master of the subject, presents a new interpretation of the evolution of the Appalachian highlands and adjacent physiographic provinces. He evolves a "theory of regional superposition" in ascribing an important rôle to a pre-Schooley peneplane that extended over the Appalachian highlands upon which is assumed to rest an extension of the coastal plain deposits across the belt of Appalachian folds. The present southeastward trending streams are presumed to have been let down from this, thus accounting for their direction. Pleasing block diagrams illustrate their development. The theory appears to be ably supported by the evidence given, and if substantiated by subsequent work will prove an enlightening hypothesis.

ALAN BATEMAN.

* Subsequent to the setting up of this review an errata slip has been received from the publisher correcting most of the errors noted.

BOOKS RECEIVED.

By

DAVID GALLAGHER.

Mineral Industry of New Jersey for 1931. By M. E. JOHNSON. Pp. 18. N. J. Dept. Conserv. and Devel., Bull. 40. Trenton, 1933. Production statistics.

Mineral Resources of Massachusetts. By I. B. CROSBY. Pp. 35. Mass. Industrial and Devel. Comm. Boston, 1932. A survey of the literature.

Anhydrite in Canada. By L. H. COLE AND R. A. ROGERS. Pp. 89, pls. 5, figs. 9. Canada Dept. Mines, Mines Branch, No. 732. Ottawa, 1933. Price, 20 cents. Properties, occurrence and utilization in Canada.

La Industria del Carbon en el Peru. By M. B. LLOSA. Pp. 111, pls. 18, maps 12. Cuerpo de Ingenieros de Minas del Peru, Boletin No. 109. Lima, 1932. Reserves, production, movements of coal, and descriptions of Peruvian mines. Maps without scales.

Brucite Deposit, Paradise Range, Nevada. By E. CALLAGHAN. Pp. 34, figs. 10. Univ. of Nevada Bull., vol. 27, No. 1. Reno, 1933. Brucite and magnesite in Triassic dolomite due to hypogene magnesian solutions from granodioritic intrusive.

Bibliography of Seismology. By E. A. HODGSON. Pp. 21. Dominion Observatory Pub., Canada, Vol. X, No. 16, Oct. Nov. Dec. 1932. Ottawa, 1933. Price, 25 cents.

State Geological Surveys and the United States Geological Survey: Summary. By M. M. LEIGHTON. Pp. 136. Nat. Res. Council Bull. 88. Washington, 1932. Price, $1. Activities, organization, appropriations, publications, past and present work, etc., of the Governmental Geological Organizations, arranged alphabetically by States.

Le Gite D'Uranium de Shinkolobwe-Kasolo (Katanga). By J. THOREAU AND R. DU TRIEU DE TERDONCK. Pp. 46, pls. 11, maps. Hayez, Bruxelles, 1933. History, mineralogy, geology, and origin of these famous uranium deposits.

Geologie Südamerikas. By H. GERTH. Pp. 199, pls. 17, figs. 38. Gebrüder Borntraeger, Berlin, 1932. One of the series "Geologie der Erde" by E. Krenkel. Part I for South America, covering the geological history to the end of the Paleozoic.

Eastern Extension of the Lupa Goldfield, Ipogolo, Sengambi and Shoga. By D. R. GRANTHAM. Pp. 9. Tanganyika Geol. Surv., Short Paper No. 11. Dar es Salaam, 1933. Price, 1/.

SCIENTIFIC NOTES AND NEWS

N. H. Darton is separated from the U. S. Geological Survey by the 30-year retirement clause in the economy act, which is being enforced in the Interior Department. This puts an end to the scientific work he has carried on for many years and will necessitate diversion into commercial or educational work.

John J. O'Neill, of McGill University, Montreal, is spending the summer as the head of a party at work on the economic geology of the eastern part of Duparquet Township, Quebec.

John A. Dresser received the honorary degree of Doctor of Laws at the annual convocation of McGill University, Montreal, on May 26.

Reginald W. Brock, dean of Applied Science, University of British Columbia, lately returned from Hong Kong, after six months spent in completing a geological survey of the island for the British Government.

K. C. Heald, geologist for the Gulf Oil Company, Pittsburgh, has been engaged in the examination of properties recently acquired by a subsidiary company, the Western Gulf Oil Company, in Fresno County, Calif.

E. T. Mellor has recently returned to England from South Africa. During his stay there he spent two months in Rhodesia as a technical witness in litigation between the Globe and Phoenix and Rhodesian Corporation mining companies.

J. B. Tyrrell, of the Royal Society of Canada, was awarded the Flavell Medal of the Society at the annual meeting at Kingston, Ontario, May 18–20.

Cyril W. Knight has been elected vice-president of Section IV (Geology) of the Royal Society of Canada.

A. Beeby Thompson, of London, recently spent several months in Iraq in the interests of British Oil Development Company, Ltd.

Lester C. Uren, of the University of California, was a guest and speaker at the annual convention of the All-Union Petroleum Scientific Society, at Baku, Russia, in June, and expected later to explore Russian and Roumanian oil fields and attend the World Petroleum Congress in London, July 19–25.

Philip D. Wilson recently spent two months in the examination of gold placer properties in Spanish Honduras and Guatemala.

Arthur Notman has been appointed by L. R. Eastman, president of

the American Arbitration Association, to serve on the business advisory council.

I. W. Jones is the leader of a party doing geological work this summer in an area south of Mont Louis, Gaspé Peninsula, Canada.

Poole Maynard, geological engineer of Atlanta, Ga., is making a laboratory investigation of the activated clays of Georgia, and states that they compare favorably with those of California and Bavaria.

The American Association for the Advancement of Science held a joint meeting with the Geological Society of America and the Geographic Society of Chicago on June 20–22 at Chicago, in connection with the Century of Progress Exposition.

The thirteenth annual meeting of the Society of Economic Geologists took place at Princeton University, Princeton, N. J., July 7 and 8. About fifty members attended, and for the first time foreign members were able to attend the meeting. The Penrose Gold Medal of the Society was presented by President B. S. Butler to L. DeLaunay of France, who received it *in absentia* through the French Consul. The following papers were presented: Calcite Marble, by G. W. Bain; Origin of Spring Hill Deposit, Helena, Mont., by Verner Jones; Auriferous Gravels of Elk and Newsome Creek, Idaho, by P. J. Shenon and J. C. Reed; Reclassification of Mineral Resources, by F. K. Morris; Iron Ores of Minas del Rif, Spanish Morocco, by A. Heim; Genesis of Fissure Veins at Przibran, by B. Stoces; Stillwater Igneous Complex, Montana, by J. W. Peoples; Magmatic Chromite in Stillwater Igneous Complex, by Edw. Sampson; Sulphides and Metamorphic Rocks at base of Stillwater Complex, by A. L. Howland; Chromite in Quebec, by B. T. Denis and J. T. Williamson; Structural Relations of Southern Appalachian Zinc Deposits, by L. W. Currier; Relation of Structure to Fixed Carbon Content of Arkansas-Oklahoma Coals, by T. A. Hendricks; Progressive Regional Metamorphism of Lower Kittanning Coal Bed of Western Pennsylvania, by T. Stadnichenko; The Molybdenum Mine of Azegour, Morocco, by A. Heim; Ore Deposits of the United States in relation to Geologic Cycles (presidential address), by B. S. Butler; Mechanism of Vein Filling at Grass Valley, California, by W. D. Johnston, Jr.; Paragenesis as an aid in Valuing New Discoveries, by B. Stoces; How to Determine Probable Changes in Primary Mineralization with Increasing Depth, by B. Stoces; Progress in Investigation of Underground Water, by O. E. Meinzer; Genetic Relations of Silver and Keweenawan Diabase in Ontario, by E. S. Moore.

V. M. Goldschmidt, professor of mineralogy at the University of Heidelberg and author of important works on crystallography, died on May 8, aged 80 years.

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